ALASKAN GLACIER STUDIES OF THE NATIONAL GEOGRAPHIC SOCIETY IN THE...

Ralph Stockman Tarr, Lawrence Martin

Mount Fairweather, 15,330 Feet High
Photographed from the Pacific Ocean by the National Geographic Society's 1909 Expedition.

ALASKAN GLACIER STUDIES

OF THE NATIONAL GEOGRAPHIC SOCIETY IN THE YAKUTAT BAY, PRINCE WILLIAM SOUND AND LOWER COPPER RIVER REGIONS

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BASED UPON THE FIELD WORK IN 1909, 1910, 1911 AND 1913
BY NATIONAL GEOGRAPHIC SOCIETY EXPEDITIONS

WASHINGTON
THE NATIONAL GEOGRAPHIC SOCIETY
1914

THE NATIONAL GEOGRAPHIC SOCIETY

The National Geographic Society, which conducted the explorations described in this volume, was organized and incorporated under the laws of the District of Columbia, January 27, 1888, for "the increase and diffusion of geographic knowledge." The Society accomplishes its object:

1. By the publication of maps, books, and an illustrated monthly magazine, which contains about 1600 pages per year. All receipts from its publications are invested in the Magazine itself or expended directly to promote geographic knowledge and the study of geography.

2. By the encouragement of geographic science and exploration by means of such financial grants as its resources will permit. The Society has just concluded a series of investigations, extending over three years, of the glaciers of Alaska, one of the most important fields of geographical research in America. The results are given in this volume. In coöperation with Yale University it has for several years maintained a large expedition in Peru, making geographical, geological and archæological investigations around Cuzco, in a region which is generally believed to have been the birthplace of the famous and little-known Inca race. It also had an expedition in Alaska investigating the recent eruption of Mount Katmai, this study being preliminary to a comprehensive investigation of what is, perhaps, the most stupendous volcanic belt on the earth.

Its earlier expeditions to Alaska did much pioneer work in the exploration of that territory. In 1902 the Society sent an expedition to Mount Pelée and La Soufrière to study the terrible eruptions of these volcanoes. The Society has assisted various Arctic expeditions, notably the last expedition of Robert E. Peary, which discovered the North Pole, April 6, 1909. In 1909 it sent to Sicily a trained geologist to investigate the Messina earthquake. A popular account of all expeditions is printed in the Magazine, while the technical results appear in separate monographs published by the Society.

3. By an annual series of addresses at the National Capital. During the past several years the Society's program has included President Taft; President Roosevelt; Secretary of State Bryan; Colonel George W. Goethals, Chief Engineer Panama Canal; Sir Harry Johnston; Sir Ernest H. Shackleton; Viscount James Bryce; President Charles W. Eliot, of Harvard University; Gifford Pinchot; Robert E. Peary; Roald Amundsen; Dr. Wilfred T. Grenfell; Sir Francis Younghusband.

4. By the maintenance of a geographic library at its headquarters in Washington.

5. By the award of gold medals.

The Society has many thousands of members distributed throughout every State in the Union, and in every foreign country. It has members in 16,000 towns and villages in the United States and in foreign lands.

The membership fee is $2 per annum with no entrance fee. Life membership fee is $50.

All members receive the *Magazine* and maps published by the Society during the term of membership.

PREFACE

The contents of this book are the result of a three-years' study of the glaciers of Alaska, financed and directed by the National Geographic Society, through its Committee on Research. The investigations were made at a cost of about $17,000, representing the interest of upward of a quarter of a million members of the Society, whose continued support of the great object of the increase and diffusion of geographic knowledge made the Alaskan undertaking possible.

Through the investigations, it is hoped, new light has been thrown upon the ice age in America, new information has been collected as to the processes of glaciation, and a new view has been gathered as to the effect of earthquakes and volcanic action upon glaciers. Alaska offered the best field in the world for these investigations, its glaciers being the largest in the world except those of the polar regions. There are thousands of them, and only a few of them even have been named.

The work of research into the extent and behavior of the Alaskan glaciers was undertaken in 1909, under a resolution of the Board of Managers appropriating $5,000 for beginning it. This was followed by like appropriations by the Board of Managers in 1910 and 1911, with some additional aid from time to time, bringing the total contribution of the Society to the work up to $17,000.

The Research Committee decided, at the outset, to devote the money expended by the Society to the study of certain of the tidewater glaciers, notably those in and about Yakutat Bay and Prince William Sound, and with the phenomena observed there this book deals.

It is peculiarly fitting that the National Geographic Society should have undertaken a scientific survey of the Alaskan glaciers, because it has been very closely identified with the exploration work that has opened up that wonderful territory. Its Magazine is one of the great storehouses of data on original glacial explorations in Alaska.

In the *National Geographic Magazine* appeared one of the first maps showing most of the Alaskan glaciers.[1] This journal published the description of C. W. Hayes and Frederick Schwatka's explorations north of the St. Elias Range and in the Copper River valley, with important contributions on glaciers and glaciation.[2] It published H. F. Reid's descriptions of the Muir Glacier and Glacier Bay.[3] It published I. C. Russell's descriptions of Mt. St. Elias and the Malaspina and Yakutat Bay glaciers,[4] first explored by the two Russell expeditions which were partly financed by the National Geographic Society. In its volumes are Henry Gannett's descriptions of some of the glaciers of Prince William Sound and other parts of Alaska[5] as well as Miss E. R. Scidmore's description of the glaciers of the Stikine River,[6] and her review of the discovery of Glacier

[1] Nat. Geog. Mag., Vol. XV, 1904, map facing p. 238.
[2] *Ibid.*, Vol. IV, 1892, pp. 117–162.
[3] *Ibid.*, Vol. IV, 1892, pp. 19–84.
[4] *Ibid.*, Vol. III, 1891, pp. 53–200.
[5] *Ibid.*, Vol. X, 1899, pp. 507–512; Vol. XII, 1901, pp. 180–196.
[6] *Ibid.*, Vol. X, 1899, pp. 5–9; Vol. VII, 1896, pp. 140–146; Vol. V, 1893, pp. 173–179.

Bay and of explorations in Alaska by members of the National Geographic Society; C. A. Stockton's [1] and J. C. Cantwell's [2] descriptions of ground ice in Arctic Alaska; C. L. Andrews' study of Muir Glacier in 1902, the first after the great earthquakes of 1899;[3] Fremont Morse's description of the same in 1907;[4] W. C. Mendenhall's descriptions of the glaciers in the Wrangell Mountains;[5] W. A. Dickey's discovery and naming of Mt. McKinley with its glaciers;[6] Robert Muldrow's determination of its height and the position of more glaciers;[7] A. H. Brooks' plan of climbing Mt. McKinley, with mapping of still more of the glaciers;[8] Ferdinand Westdahl's photographs of the glaciers and snowfields on the volcanoes of Unimak Island in the Aleutians;[9] W. H. Osgood's description of the glaciers near Lake Clark;[10] G. K. Gilbert's summary of the glaciers of Alaska;[11] and many other papers[12] dealing incidentally with glaciers and glaciation in Alaska.

The National Geographic Society's Alaskan expedition of 1909, the plan of which was announced in the *Magazine* for June, 1909, was under charge of the authors of this volume, the other members of the party being W. B. Lewis of the U. S. Geological Survey, the topographer; Oscar von Engeln of Cornell University, who took and developed many of the photographs; E. F. Bean of the University of Wisconsin, who acted as rodman and general assistant; A. R. Campbell of the University of Washington, who ran the launch; Charles Johnson of Yakutat, who was boatman and camp hand; and a Japanese boy, as cook.

The Society's 1910 expedition was under the leadership of the junior author, assisted by Messrs. Lewis and Bean of the 1909 party, F. E. Williams of the University of Wisconsin, rodman; R. G. Byers and E. A. Conner of the University of Washington, photographer and boat engineer respectively; and a cook.

The Society's 1911 expedition, under the direction of both authors of this volume, devoted most of the season to glacier study in other parts of Alaska than are described in this volume. The junior author, however, gave two weeks to additional observations of the Prince William Sound and Copper River glaciers before he was joined by the senior author and Russell S. Tarr.

The junior author also visited Yakutat Bay in 1913.

Each of the authors had previously spent two summers in Alaska, the senior author in Yakutat Bay in 1905 and 1906, the junior author in Controller Bay, Prince William Sound, Cook Inlet, and Alaska Peninsula in 1904, and in Yakutat Bay in 1905. Both of these previous seasons of field work in Alaska by the senior author were supported by the U. S. Geological Survey. The 1904 season by the junior author was in a U. S. Geological Survey party, but in 1905 he was supported by a grant of money from the American Geographical Society of New York.

[1] Nat. Geog. Mag., Vol. II, 1890, pp. 178–179.
[2] *Ibid.*, Vol. VII, 1896, pp. 345–346.
[3] *Ibid.*, Vol. XIV, 1903, pp. 441–445.
[4] *Ibid.*, Vol. XIX, 1908, pp. 76–78.
[5] *Ibid.*, Vol. XIV, 1903, pp. 395–407.
[6] *Ibid.*, Vol. VIII, 1897, pp. 322–327.
[7] *Ibid.*, Vol. XII, 1901, pp. 312–313.
[8] *Ibid.*, Vol. XIV, 1903, pp. 30–35.
[9] *Ibid.*, Vol. XIV, 1903, pp. 93, 94, 97, 98.
[10] *Ibid.*, Vol. XV, 1904, p. 328.
[11] *Ibid.*, Vol. XV, 1904, pp. 449–450.
[12] *Ibid.*, Vol. XX, 1909, pp. 585–623; Vol. XXII, 1911, pp. 597–600, 786; Vol. XXIII, 1912, pp. 428–429, 684–713.

We sailed from Seattle June 24, 1909, on the Alaska Coast Co.'s steamer, *Portland*, reaching Yakutat five days later. Here we spent about six weeks. We traveled and transferred our food and camp equipment in a twenty-eight foot whaleboat, equipped with a four-horse-power gasoline engine. We occupied two main camps and seven temporary camps in Yakutat Bay. We left Yakutat on August 14th, taking our launch upon the steamer from Yakutat to Valdez. Two days apiece were devoted to brief examinations of Valdez, Shoup, and Columbia glaciers, in eastern Prince William Sound, traveling in our launch, and camps being occupied at Fort Liscum opposite Valdez, on Heather Island, near Columbia Glacier, and on Flemming's Spit just outside Cordova. Two days were devoted to a hurried view of the Miles, Childs, and Allen glaciers of lower Copper River, the authors being taken by railway automobile to the end of the line, as then completed, and entertained at the railway construction camps. We left Cordova September 1 on the Northern Navigation Company's steamer *Lindsay*, reaching Seattle ten days later.

During the field season of 1910, the junior author and his party spent only two weeks in Yakutat Bay, June 11 to 26, occupying five camps and traveling in the launch, which had been purchased the previous year by the Society. The party was then taken to Prince William Sound, where a special trip of Alaska Coast Company's steamer *Bertha* landed us on Heather Island near Columbia Glacier. In the following six weeks, we traveled by launch clear around Prince William Sound, reaching Cordova on August 12, after occupying ten camps in the fiords near the several glaciers and on Latouche and Hinchinbrook Islands. Three weeks were then devoted to the glaciers of the lower Copper River. We worked as far north as Heney Glacier and had four camps along the Copper River and Northwestern Railway near the larger glaciers. The junior author also spent three days in a rapid railway and steamboat trip through the Copper River canyon to Chitina. A week was occupied in a launch trip from Cordova to Columbia Glacier and Valdez, whence the party sailed for Seattle on the Alaska Steamship Company's steamer *Northwestern* on September 9.

In 1911 the junior author went north alone, leaving Seattle June 5 on the Alaska Steamship Company's steamer *Alameda*, making a rapid reconnoissance trip to Valdez Glacier and through Prince William Sound to Seward and back to Cordova, after which he spent the time from June 15 to 20 at the glaciers of the lower Copper River. He was joined by the senior author and after another trip through Prince William Sound and to Columbia Glacier and a study of glaciers in the Kenai Peninsula along the Alaska Northern Railway, we went through the Copper River canyon by train, saw something of the conditions of glaciation in a day's work at Wood Canyon, and then worked the remainder of the season on the glaciers and glaciation of the interior and of southeastern Alaska.

In 1913 the junior author was guide for an excursion of the International Geological Congress which visited Yakutat Bay in September. The party consisted of about 50 geologists from all parts of the world. We travelled and lived on the specially-chartered Canadian Pacific steamer *Princess Maquinna*, and landed at a number of points, as described in later pages of this book.

Preliminary reports of the results obtained in 1909 and 1910 were published in the *National Geographic Magazine* for January, 1910, and June, 1911. The junior author lectured before the Society in Washington on February 18, 1910.

The phenomena of advancing glaciers observed in Yakutat Bay by the U. S. Geological

Survey parties in 1905 and 1906 were further studied by the subsequent expeditions of the National Geographic Society. The fullness of the description of the Yakutat Bay region, therefore, depends not merely on the work of the seasons from 1909 to 1913, but to a very large degree, on the studies of 1905 and 1906. The investigations of these five seasons have developed a series of interesting and important results, the statement of which can be adequately made only by making free use of the work of the previous expeditions. This has, therefore, been done, not in the form of direct quotation of our previous publications relating to the expeditions of 1905 and 1906, but in such abstracts as have seemed necessary to the clearness of discussion of the problems which the Yakutat Bay glaciers present. This report, in so far as the Yakutat Bay region is concerned, is the summary and discussion of the results of five seasons of work, not of the seasons of 1909, 1910 and 1913 alone.

The description of conditions in Prince William Sound and on the lower Copper River is almost entirely new. The discussion of the glaciers of Alaska in Chapter I of this book, and of the glaciers and glaciation of Prince William Sound and the Lower Copper River, Chapters XII to XXIII, is wholly the work of the junior author. The senior author, however, with his greater experience and wide observation of glaciers in Greenland, Norway, Spitzbergen, and the Alps, saw a great deal in the few days he spent in eastern Prince William Sound and on the lower Copper River in 1909. Accordingly he was able to advise in the arrangement of the materials of Chapters XII to XXIII and generously gave much time and thought to the criticism of the portions of manuscript—all but the last three chapters—which were completed before his lamentable death on March 21, 1912. He also wrote parts of the pages on the Comparison between Yakutat Bay and Prince William Sound. The junior author's debt to Professor Tarr, as teacher in lecture room, laboratory, and field, and as councillor and friend can never be adequately acknowledged.

Many of the photographs used as illustrations in this book were taken by the authors and by our photographers—O. D. von Engeln in 1909, and R. B. Byers in 1910. We are deeply indebted to both of them for their excellent photographs and for splendid service in the field in Alaska, where the exposing and developing of photographs have especial difficulties. In the *Magazine* for January, 1910, the former of these men described some of these difficulties. We have also used a few photographs taken by the late Professor I. C. Russell of the University of Michigan and Mr. H. G. Bryant of Philadelphia, by members of the Harriman Expedition, of the United States Geological Survey, and the International Boundary Surveys, by engineers of the Copper River and Northwestern Railway, and several Alaskan photographers, as acknowledged specifically in the legends of the plates.

Among the text figures are several based upon manuscript maps of Prince William Sound, supplied by Professor U. S. Grant of Northwestern University. Some of the cross-sections of the fiords were drawn by Mr. E. F. Bean of the University of Wisconsin, who has also been of great service in assisting with the proofreading and with making the index of this book.

We were thoroughly equipped for glacier study during each season in the field, thanks to the large appropriations by the Research Committee. For assistance in arranging and carrying out our plans in Alaska we are under deep obligations to the companies and institutions whose contributions are acknowledged elsewhere in this book, as well

as to the following gentlemen: Rev. Mr. Rasmusson, and Messrs. Beasley, Stimson, Robinson, Flenner, and Gray, of Yakutat; the late E. C. Hawkins, General Manager, and Mr. S. W. Eccles, the President of the Copper River & Northwestern Railway Co.; the late Alfred Williams and Messrs. O'Neel, Van Cleve, Johnson, Withers, Corser, and Wernicke of the Katalla Company, and Messrs. Murchison, Shields, McCune and Whiting of M. J. Heney's force; Messrs. Dalton and Hazelet of Cordova; the late Dr. L. S. Camicia and Messrs. Lathrop, Peterson, and Crawford of Valdez; Captain MacGilvary and Messrs. Johansen and Inglis of the Steamship *Bertha*, and Messrs. Barber and Macgregor of the steamship *Portland;* the mine operators at Latouche; Professors Landes and Meany of the University of Washington; the late Webster Brown and Messrs. George Jamme and J. L. McPherson of Seattle, and R. P. Tarr of Tacoma; Commander Ellicott of the U. S. Navy; Capt. C. G. Quillian of the U. S. Coast and Geodetic Survey; Col. A. W. Swanitz, Chief Engineer, and Mr. O. G. Laberee, President of the Alaska Northern Railway; Col. Richardson of the Alaska Road Commission; Mr. A. B. Emery, Manager of the Bonanza copper mine at Kennicott; Messrs. James, Fisher, Rhodes, Bigelow, Coleman, and Watson of Fairbanks and Manager Joint of the Tanana Valley Railway; Messrs. Perry and Coffee of The Yukon Gold Company at Dawson, Mr. Cobb and Judge Gunnison of Juneau; Mr. Bronson, U. S. Collector of Customs at Wrangell; Mr. George Otis Smith, Director of U. S. Geological Survey; Mr. A. H. Brooks, geologist in charge of Division of Alaskan Mineral Resources of the U. S. Geological Survey; Mr. O. H. Tittmann, Superintendent of the U. S. Coast and Geodetic Survey; Mr. W. F. King of the Dominion Astronomical Observatory at Ottawa; Messrs. Fremont Morse and N. J. Ogilvie of the International Boundary Surveys; Prof. H. F. Reid of Johns Hopkins University, and many others.

We wish to acknowledge the generous support of every member of our parties in the field, the cordial hospitality of all who met us in Alaska, and the assistance of the officers of the Society, particularly Mr. Gannett, the President of the Society and Chairman of the Research Committee, Mr. Grosvenor, the Editor of the *National Geographic Magazine* and the Director of the Society, and the other members of the Research Committee, in doing everything in their power to make our field seasons effective and in facilitating the preparation and publication of this book.

LAWRENCE MARTIN.

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ALASKAN GLACIER STUDIES

CHAPTER I

THE GLACIERS OF ALASKA

The Alaskan glaciers are the largest in the world except those in polar regions.

The existing glaciers of Alaska are chiefly confined to the mountain regions and extend more or less continuously through about 1,100 miles along the Coast and the St. Elias Ranges, 400 miles in the Alaska Range, and 450 miles in the Endicott Mountains of northern Alaska. These mountain, snowfield, and glacier belts average from 40 to 120 miles in width. There are also scattered glaciers upon volcanoes through a distance of 500 miles on the Alaska Peninsula and Aleutian Islands. Glaciers are found through a range of 15 degrees of latitude. The total snow- and ice-covered area has been conservatively estimated at 15,000 to 20,000 square miles,[1] and is much smaller than the area covered at the maximum stage of Alaskan glaciation. This, however, is less than four per cent of the total area of Alaska. There are thousands of glaciers, only a few hundred of which have been named.

The glacial districts in Alaska may be best described in groups associated with the various mountain ranges (Map 1, in pocket), both because these enable us to take up compact units, and because the mountains are jointly responsible with the climatic conditions for the existing glaciation. It is the combination of lofty mountains facing a sea coast where warm, humid, onshore winds bring abundant moisture, in a northerly latitude, that gives the Pacific Mountains of Alaska from 80 to 200 inches of precipitation yearly. It is the loftiness of these mountains, and the northerly latitude, that causes a large proportion of this precipitation to fall in the form of snow. Therefore, much more snow falls in a winter than can melt during a summer, causing permanent snowfields and great glaciers. The variations in altitude, in latitude, in precipitation, and in direction of slope cause the principal variations in the present size and condition of the glaciers. Most of these variations are associated with differences in the mountains, but it is not certain that the climate is responsible for all the glacial oscillations, nor that the ice tongues are consistently waning.

The mountain ranges upon whose slopes most of the glaciers have been formed are the Pacific Mountains, including the various subdivisions of (1) the Canadian Coast Range, (2) the St. Elias Range, (3) the Alaska-Aleutian Ranges.

Former glaciation will be discussed in connection with all these mountain groups after the existing glaciation has been sketched.

Along the Inside Passage the Taku and Davidson Glaciers are rather familiar, as are the Muir Glacier of Glacier Bay, the Malaspina and Yakutat Bay Glaciers of the south side of the St. Elias Range, and the Valdez and Columbia Glaciers of Prince William Sound; but most of the other glaciers of Alaska are little known.

[1] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, p. 9; Nat. Geog. Mag., Vol. XV, 1904, p. 450.

Brief general lists or descriptions of the glaciers of Alaska have previously been made by Muir,[1] Petroff,[2] Russell,[3] Dall,[4] Wright,[5] Baker,[6] Davidson,[7] Brooks,[8] Gilbert,[9] Greely,[10] and, for individual glaciers, by others listed upon subsequent pages. The glaciers of the whole territory were first shown upon a map by R. U. Goode and E. C. Barnard, published in the National Geographic Magazine [11] and afterwards reproduced by the United States Geological Survey.[12]

Canadian Coast Range[13]

Glaciers near Portland and Behm Canals. Beginning with the panhandle of southeastern Alaska the first existing glaciers north of British Columbia are those of the Bear and Salmon River headwaters of Portland Canal, and the Chickamin, Leduc, and Unuk River tributaries of Behm Canal. Here fifty or sixty valley glaciers, the larger ones 2 to 10 miles long and 1 to 2½ miles wide, descend from snowfields at an elevation of four to eight thousand feet and end at elevations ranging from 500 to 3000 feet above sea level. None of these ice tongues, except the Soulè Glacier [14] as yet bears a name. Photographs showing some of them were taken by the Canadian Boundary Commission in 1894, and these photographs, as well as the detailed maps showing the glaciers,[15] are of much value because a second series of photographs and maps made by the commission engaged in locating the boundary in 1906 to 1913 will show the advances and retreats of these ice tongues after more than ten years. The maps unfortunately are on a small scale (1:160,000). They do not go far back into the snowfield area, nor do they show all the glaciers, particularly those back in the interior of the Coast Range. Glaciology has, however, a great aid here in the two sets of maps made as a result of the

[1] Muir, John, In newspaper notices even before Petroff, Russell, and Dall; Notes on the Pacific Coast Glaciers, American Geologist, Vol. XI, 1893, pp. 287–299; Harriman Alaska Expedition, Vol. I, 1901, pp. 119–135.

[2] Petroff, Ivan, Tenth Census of the United States, 1880, Vol. VIII, 1884, pp. 24, 85–86, and Map VI facing p. 75.

[3] Russell, I. C., Fifth Ann. Rept., U. S. Geol. Survey, 1883–4, pp. 348–355; Glaciers of North America, Boston, 1897, pp. 74–130; Scottish Geog. Mag., Vol. X, 1894, pp. 405–407 and map showing known glaciers.

[4] Dall, W. H., Glaciation in Alaska, Bull. Phil. Soc., Washington, Vol. 6, 1884, pp. 33–36.

[5] Wright, G. F., Ice Age in North America, New York, 1891, 13–35.

[6] Baker, Marcus, Geographic Dictionary of Alaska, Bull. 187, U. S. Geol. Survey, 1902, 2d edition; Bull. 299, 1906.

[7] Davidson, George, The Glaciers of Alaska that are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc. Pacific, 2d series, Vol. 3, 1904, pp. 1–98.

[8] Brooks, A. H., Glacial Phenomena of Southeastern Alaska, in Professional Paper 1, U. S. Geol. Survey, 1902, pp. 31–35; Geography and Geology of Alaska, Professional Paper 45, U. S. Geol. Survey, 1906, pp. 244–248, 295–296, Pls. XII and XXII.

[9] Gilbert, G. K., Glaciers, Harriman Alaska Expedition, Vol. III, New York, 1904, pp. 1–223; Nat. Geog. Mag., Vol. XV, 1904, pp. 449–450.

[10] Greely, A. W., Handbook of Alaska, 1909, pp. 152–159, 265–267.

[11] Brooks, A. H., Nat. Geog. Mag., Vol. XV, 1904, opposite p. 238.

[12] Professional Paper 45, U. S. Geol. Survey, 1906, Pl. XXXIV; Bull. 345, Pl. I, 1908; Map A, Alaska, U. S. Geol. Survey, 1909.

[13] Some of these glaciers are shown on Chart 8050, U. S. Coast and Geod. Survey, Dixon Entrance to Head of Lynn Canal; also on the map of Southeastern Alaska and a part of British Columbia, Showing Award of Alaskan Boundary Tribunal.

[14] Chart 8100, U. S. Coast and Geod. Survey.

[15] Sheets 4 and 7, Atlas of Award, Alaskan Boundary Tribunal, Senate Document 162, 58th Congress, 2d Session, Washington, Government Printing Office, 1904.

surveying necessary to the boundary arbitration. F. E. and C. W. Wright have briefly alluded to these glaciers and the wonderful glacial sculpture of the adjacent fiords.[1]

In the valley of Blue River, and adjacent northwest tributaries of Unuk River is an interesting series of recent lava flows, the latest thought to be less than 50 years old, described by Morse of the Boundary Survey. One or more of the small glaciers near Blue River is covered with ashes.[2]

Glaciers of Stikine River. On the lower Stikine River and partly in Canada are the Great Glacier, the Popoff or Little Glacier, the Dirt or Mud Glacier, the Flood Glacier, and many others shown on the Boundary Atlas Sheets [3] already referred to, and on one Coast Survey chart [4] and one U. S. Geological Survey map.[5] The Stikine River glaciers were visited and briefly described by Blake in 1867,[6] Hunter in 1877,[7] by Muir and by Bell in 1879,[8] by Dawson in 1887,[9] by Miss Scidmore in 1898,[10] and others. Great Glacier is an ice tongue of which only the lower sixteen miles have been mapped. The upper glacier is over 4 miles wide, the lower valley about a mile wide, and beyond this the Great Glacier spreads out to a width of four and a half miles in a piedmont bulb which enters the Stikine valley from the west, forcing the Stikine River over to the east bank. The terminus of this first large glacier in Alaska is less than 250 feet above sea level. The Dirt Glacier seems to be stagnant and moraine-covered. Russell shows photographs of Orlebar and Bernard Glaciers on the Stikine River,[11] which Dawson states are the Great Glacier and either the Flood or Dirt Glaciers respectively. The débris carried out from these glaciers has caused the Stikine River to build a great delta seventeen miles wide which nearly ties Mitkof Island to the mainland and makes it necessary for ships to go through the dangerous Wrangell Narrows in navigating the Inside Passage.

Glaciers of Frederick Sound and Stephens Passage. This group of ice tongues includes the LeConte, Patterson, Baird, Dawes, Brown, Sumdum and Sawyer Glaciers and the unnamed glaciers of Whiting and Speel Rivers. They are shown on the Coast Survey [12] and Boundary Survey [13] and Geological Survey [14] maps. The LeConte Glacier in latitude 57° is the southernmost tidal glacier in Alaska. The Dawes, Brown, and Sawyer Glaciers also reach sea level and discharge icebergs.[15] The group of glaciers near Devil's

[1] Bull. 347, U. S. Geol. Survey, 1908, p. 24 and map, Pl. II.

[2] Morse, Fremont, Nat. Geog. Mag., Vol. 17, 1906, pp. 173, 176.

[3] Atlas of Award, Alaskan Boundary Tribunal, Sheets 8 and 9.

[4] Chart 8200, U. S. Coast and Geod. Survey, 1891.

[5] Bull. 347, U. S. Geol. Survey, 1908, Pl. III.

[6] Blake, W. P., The Glaciers of Alaska, Russian America, Amer. Journ. Sci., 2d series, Vol. XLIV, 1867, 96–101; House Extra Doc., 177, Part II, 40th Congress, 2d Session, with map.

[7] Hunter, J., Map reproduced in Alaskan Boundary Tribunal, Maps and Charts accompanying the case of Great Britain, Washington, 1904, p. 26.

[8] Bell, W. H., Scribner's Monthly, Vol. XVII, 1879, pp. 805–815.

[9] Dawson, G. M., Geol. and Nat. Hist. Survey Canada, Ann. Rept., Vol. III, Part I, 1889, pp. 51B–53B.

[10] Scidmore, E. R., Nat. Geog. Mag., Vol. X, 1899, pp. 5–9.

[11] Russell, I. C., 5th Ann. Rept., U. S. Geol. Survey, 1883–1884, Pl. LIV and LV.

[12] U. S. Coast and Geod. Survey, Charts 8200, 8210, 8300.

[13] Atlas of Award, Alaskan Boundary Tribunal, Sheets 9, 10, 11, 12.

[14] Bull. 287, U. S. Geol. Survey, 1906, Pl. XXXVI; Bull. 347, 1908, Pl. III.

[15] See description by Sir George Simpson in September, 1841, Journey Round the World, Vol. I, p. 213, quoted by Klotz, Geog. Journ., Vol. XIV, 1899, pp. 531–532.

Thumb are the southernmost Alaskan glaciers seen from the steamer, Patterson Glacier being visible from the north end of Wrangell Narrows.

The increase in glaciation with increase in latitude may be shown by the lower snowline on Mt. Sumdum, than on the mountains to the south, as well as by the presence of tidal glaciers. Larger glaciers appear now, the shorter arm of Baird Glacier being sixteen miles long, the longer tributary being certainly over 17 miles; but its whole length has not been mapped. They advance from vast snowfields whose extent is not yet known. Some, notably the Dawes and Baird, are *through glaciers*, or ice tongues which flow in opposite directions from a common, flattish, snow-covered divide. Some of the glaciers of this region bifurcate so close to their terminus that one name is used for both arms, as in Baird Glacier. The name Dawes is used for two adjacent glaciers which formerly were one. The same is true of the Sawyer Glaciers.

The extreme glacial erosion and the lack of glacial deposits in southern Frederick Sound has been pointed out by the Wrights.[1]

Klotz determined that LeConte Glacier retreated a half mile between the time of making a Coast Survey chart, probably in the late eighties, and 1893.[2] The bay it enters is called Hutli or Thunder Bay from the noise made by the discharge of icebergs, according to John Muir who visited this glacier in 1879.[3] Icebergs then filled the bay for ten or twelve miles. Patterson Glacier was advancing and destroying trees in 1891 according to the Pacific Coast Pilot of that year.[4] Baird Glacier, not to be confused with the other Baird Glaciers of the Tasnuna River and White Pass, has been studied by Klotz,[5] who determined its rate of movement and rate of melting in 1894.

Holkham Bay, at the base of Mt. Sumdum, with its branches Endicott and Tracy Arms, is said by Muir[6] to be one of the most interesting Alaskan fiords. It has four tidal glaciers and a hundred or more glaciers of the second and third class. The icebergs interfere with navigation by large boats and it is often difficult to reach the tidal glaciers in small canoes.

The unnamed ice tongues of Speel and Whiting Rivers are exceedingly attenuated, narrow glaciers, 6 or 7 miles long and a tenth to a quarter of a mile wide.

Glaciers of Taku Inlet. This beautiful fiord just south of Juneau, is visited annually by thousands of tourists and the Taku Glacier has become one of the best known in Alaska, especially during the past 15 years when Muir Glacier has been less accessible and has lost beauty through retreat and loss of height. Within the fiord are also the Norris, Wright, and Twin Glaciers, besides a few smaller glaciers up the Taku River,[7] but the Taku Glacier is the only one discharging icebergs. These glaciers are shown on several maps [8] and there are also innumerable smaller glaciers. Muir says forty-five glaciers are visible from a steamer sailing up the middle of Taku Inlet.[9]

[1] Wright, F. E. and C. W., Bull. 347, U. S. Geol. Survey, 1908, p. 24 and Pl. III.

[2] Klotz, Otto, Geog. Journ., Vol. 14, 1899, p. 532.

[3] Muir, John, Amer. Geol., Vol. XI, 1893, p. 291.

[4] Reid, H. F., Variations of Glaciers, Journ. Geol., Vol. VIII, 1900, p. 158.

[5] Klotz, Otto, Journ. Geol., Vol. III, 1895, pp. 512–518.

[6] Muir, John, Harriman Alaska Expedition, Vol. I, 1901, p. 125.

[7] Hayes, C. W., Nat. Geog. Mag., Vol. IV, 1892, p. 151.

[8] U. S. Coast and Geod. Survey, Charts 8050, 8300; Atlas of Award, Alaskan Boundary Tribunal, Sheets 12 and 13; Pl. 19, Vol. IV, Nat. Geog. Mag., 1892; Pl. XXXVI, Bull. 287, U. S. Geol. Survey, 1906.

[9] Muir, John, Harriman Alaskan Expedition, Vol. I, 1901, p. 125.

Taku Glacier for which the names Schulze Glacier and Foster Glacier were temporarily used, is 30 miles long, heading to the north on a 5,500-foot snow divide as a through glacier, the other end of which flows down the east side of the Canadian Coast Range. This may be the Llewellyn Glacier[1] of Atlin Lake. To the northwest it shares snowfields as a through glacier with Mendenhall and Herbert Glaciers of Lynn Canal and Norris Glacier of Taku Inlet. Taku Glacier receives at least ten tributaries, from a half mile to a mile and a half wide, the main glacier being two to three miles wide. It gives off two glacial distributaries, one of which feeds the west half of the Twin Glaciers of Taku Inlet. The sea front is a little over a mile wide and 200 feet high and has been described by Vancouver,[2] Muir,[3] Russell,[4] and many others.[5] Taku Glacier suffered severely during the Yakutat Bay earthquakes of September, 1899, but had a net advance between 1890 and 1905.[6] Norris Glacier, to which the name Windom Glacier was temporarily applied, is reported [7] to have lost part of its end following the 1899 earthquakes, perhaps because of washing away of supporting gravels.

Glaciers East of Lynn Canal. On the east side of this great fiord a group of glaciers descend from the snowfields of the Canadian Coast Range, the better-known valley glaciers, named in order from Juneau northward to Skagway, being the Mendenhall,[8] Herbert, Eagle, Meade,[9] and Ferebee Glaciers. They are well shown on a number of maps.[10]

Besides these large glaciers that bear names there are as many glaciers of equal size that are as yet unnamed because not seen from Lynn Canal, as for example the five glaciers terminating ten to fifteen miles back from Lynn Canal in the valleys draining to Berners Bay. There are also hundreds of minor glaciers, such as the Lemon Creek Glacier and others near Juneau,[11] and the glaciers on the north side of Lion's Head Mountain, Berners Bay.[12]

The region from which the Mendenhall, Taku, and adjacent glaciers extend is a typical ice-flooded region, an area of many square miles between Taku Inlet, Lynn Canal, and the international boundary having, according to the map, a larger proportion of snowfields and valley glaciers than of projecting mountain ridges and isolated peaks. Many of the ice tongues are through glaciers.

[1] Gwillim, J. C., Geol. Survey of Canada, Ann. Rept., Vol. XII, 1902, pp. 14B–15B and map 742.

[2] Quoted by Klotz, Geog. Journ, Vol. XIV, 1899, p. 531.

[3] Muir, John, Amer. Geol., XI, 1893, p. 293.

[4] Russell, I. C., Glaciers of North America, Boston, 1897, pp. 78–80.

[5] For example, Davidson, George, Trans. and Proc. Geog. Soc. Pacific, Vol. III, 1904, pp. 79–81.
Egerton, H. G., Alaska and Its Glaciers, Nineteenth Century, Vol. 32, p. 1001.
Higginson, Ella, Alaska, the Great Country, New York, 1908, pp. 110–113.
Greely, A. W., Handbook of Alaska, New York, 1909, pp. 153–154.

[6] Reid, H. F., Variations of Glaciers, Journ. Geol., Vol. XIII, 1905, p. 317; Same, Vol. XIV, 1906, p. 408.

[7] Reid, H. F., Variations of Glaciers, Journ. Geol., Vol. XI, 1901, p. 253.

[8] Knopf, A., Bull. 502, U. S. Geol. Survey, 1912, pp. 11–13, 32–33, and Pl. I.

[9] Sheldon, Charles, The Wilderness of the Upper Yukon, New York, 1911, pp. 172–174.

[10] Coast Survey Charts 8050, 8300, 8302, 8303.
Atlas of Award, Alaskan Boundary Tribunal, Charts 13, 16a, 17.
Bull. 287, U. S. Geol. Survey, 1906, Pl. XXXVIII.

[11] Eldridge, G. H., Maps and Descriptions of Routes of Exploration in Alaska in 1898, U. S. Geol. Survey, Washington, 1899, p. 102.
Juneau Special Map, 1904, U. S. Geol. Survey.

[12] Knopf, A., Bull. 446, U. S. Geol. Survey, 1911, p. 12 and Pl. I; Berners Bay Special Map, Alaska, Sheet No. 581, 1906, U. S. Geol. Survey.

None of these glaciers are now tidal, although Mendenhall Glacier, which is 17 miles long and two to three miles wide at its terminus, descends to within less than 100 feet of sea level. References to floating ice here in 1794 by Vancouver's lieutenant, Whidbey, and by Sir George Simpson in 1841, as well as the Admiralty and Hydrographic charts of 1865 and 1869, have convinced Davidson that Mendenhall Glacier reached tidewater in those years.[1] He quotes Manson's observations, based on a miner's stakes near Mendenhall Glacier, as proving a retreat of 40 or 50 feet a year between 1892 and 1901.

Davidson has briefly described the glaciers and glaciation of the head of Lynn Canal or Taiya Inlet[2] and Schwatka[3] has described the Saussure, Baird, and several other small glaciers between Lynn Canal and Lake Lindemann. Andrews reports that eight glaciers near Skagway were all retreating rapidly in 1903, one having gone back 30 or 40 feet annually since 1898.[4] The Krause brothers visited some of these glaciers in 1881.[5] Denver Glacier is visited annually by hundreds of tourists.

White Pass and Chilkoot Pass, which were crossed by thousands of gold-seeking prospectors in the rush to the Klondike during the late nineties, are not covered by glaciers, although there are many ice tongues in the mountains nearby. These passes, 2,600 and 3,600 feet respectively above sea level, were snow-covered, however, at the time of the rushes, especially in the early spring. Russell has called attention to the 5 or 6 small glaciers seen by him in 1889, and commented upon the fact that those on the north side of the Coast Range are much smaller than those on the south[6] where he saw forty small glaciers from a single viewpoint. In 1896 J. E. Spurr observed that the glaciers on the south side of the Coast Range were retreating.[7]

St. Elias Range

Glaciers West of Lynn Canal. On the west side of Lynn Canal a series of glaciers descend from the slopes of the St. Elias Range, though none reach sea level.[8] These include a number of valley glaciers in the region drained by the western tributaries of the Chilkat River, including the Knapp, Leslie, Jarvis, Le Blondeau, Takhin, Bertha, and Garrison Glaciers and the glaciers whose streams flow directly to Chilkat Inlet or Lynn Canal, such as Rainbow Glacier, the well-known piedmont bulb of Davidson Glacier, and a series of unnamed ice tongues, some of which are through glaciers descending from the snowfields that also feed Muir Glacier on the west side of the same range. Wright has discussed the general glaciation of the Porcupine gold district.[9]

[1] Davidson, George, The Glaciers of Alaska that are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc. Pacific, Vol. III, 1904, pp. 78–79.

[2] Bull. Phila. Geog. Soc., Vol. II, 1900, pp. 108–114.

[3] Science, Vol. III, 1884, p. 220; Report of a Military Reconnaissance Made in Alaska in 1883, Compilation of Narratives of Exploration in Alaska, Washington, 1900, pp. 295–296 and map.

[4] Quoted by Reid, H. F., Variations of Glaciers, Journ. Geol., Vol. XII, 1904, p. 260.

[5] Krause, Arthur and Aurel, Ergebnisse einer Reise nach der Nordwest Küste von Amerika und der Berings-Strasse, Jena, 1885; Krause, Arthur, Zeitschrift der Ges. für Erdkunde zu Berlin, Vol. XVIII, 1883.

[6] Russell, I. C., Surface Geology of Alaska, Bull. Geol. Soc. Amer., Vol. I, 1890, pp. 148–151.

[7] Quoted by Reid, H. F., Variations of Glaciers, Journ. Geol., Vol. V, 1897, p. 381.

[8] Coast Survey Charts 8050, 8300, 8302, 8303.

Atlas of Award, Alaskan Boundary Tribunal, Sheets 14, 17, 18.

Pl. XLIX, 21st Ann. Rept., U. S. Geol. Survey, Part II, 1899–1900; Pl. XXXVII; Bull. 287, U. S. Geol. Survey, 1906.

[9] Wright, C. W., Bull., 236, U. S. Geol. Survey, 1904, pp. 14, 18–19, and Pl. II, V.

In this region the Takhin and Le Blondeau Glaciers are of especial interest because they seem to have blocked the Takhin valley, about 25 miles west of Haines Mission, diverting the upper ten miles of this stream northeastward into Salmon River or Tsirku Creek which flows into Chilkat River at Klukwan.[1] Rainy Hollow Glacier had an advance of over 2000 feet between June and September, 1910, as observed by Webster Brown.[2]

Davidson Glacier which has been described by Davidson, Blake, Muir, Meehan, Russell, Gilbert, and others [3] seems to have been one of the first bulb glaciers recognized in Alaska, Russell having described and explained its fan-shaped terminus.

Glaciers of Glacier Bay. These, the most visited glaciers in Alaska, include the Muir, Carroll, and Rendu ice tongues that flow south or west from the portion of the St. Elias Range just described, the Wood, Geikie, Charpentier, Hugh Miller, and Johns Hopkins Glaciers that flow eastward from the Fairweather Range, and the Grand Pacific Glacier which flows southward between these two divisions of the St. Elias Range, heading as a through glacier, as does the northern part of the Muir, on a flat divide that sends an ice tongue northward to the Alsek River. There are eleven tidal glaciers here, of which Muir Glacier alone has a drainage area of over 800 square miles and over 350 square miles of glacier surface. Its two main tributaries are 20 and 22 miles long.

Glacier Bay was visited by Vancouver in 1794 and later by several of the Russian explorers. The modern visits begin with those of Lieutenant Wood in 1877 [4] and John Muir in 1879 and 1880.[5] Glacier Bay was discovered in the sense of being made known to the world [6] by John Muir, whose name is fittingly attached to the grandest of its glaciers. After Muir's visits, Muir Glacier was first studied scientifically by G. W. Lamplugh in 1884 [7] and by G. F. Wright in 1886.[8] All the ice tongues were studied and mapped carefully by H. F. Reid in 1890 and 1892,[9] H. P. Cushing sharing the observations of Muir Glacier in 1890.[10] I. C. Russell also made brief observations of Muir Glacier in 1890.[11] The region was remapped by the Canadian Boundary Survey in 1895,[12] and revisited by Muir in 1896.[13]

[1] See Pl. XXXVII, Bull. 287, U. S. Geol. Survey.

[2] Journ. Geol., Vol. XXI, 1913, p. 426.

[3] Davidson, George, House Extra Doc. 177, 40th Congress, 2d session, 1868, p. 276; Glaciers of Alaska, etc., Trans. and Proc. Geog. Soc. Pacific, Vol. III, 1904, pp. 73–76.

Blake, T. A., House Ex. Doc., 40th Congress, 2d Session, 1868, p. 321.

Muir, John, Amer. Geol., Vol. XI, 1893, p. 293.

Meehan, Thomas, Notes on Glaciers in Alaska, Proc. Philadelphia Acad. Science, 1883, pp. 249–255.

Russell, I. C., Bull. Geol. Soc. Amer., Vol. I, 1890, p. 152.

Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 12–16.

[4] Wood, C. E. S., Century, Vol. II, 1882, pp. 333–335.

[5] Muir, John, Amer. Geol., Vol. XI, 1893, pp. 294–299; Century, Vol. XXVIII, 1895, pp. 234–247; Harriman Alaska Expedition, Vol. 1, 1902, pp. 125–128.

[6] Scidmore, E. R., The Discovery of Glacier Bay, Alaska, Nat. Geog. Mag., Vol. VII, 1896, pp. 140–146.

[7] Nature, Vol. 33, 1886, pp. 299–301.

[8] Wright, G. F., Amer. Journ. Sci., Vol. XXXIII, 1887, pp. 1–18; Ice Age in North America, 1891, pp. 36–66; Man and the Glacial Period, 1892, pp. 24–30.

[9] Reid, H. F., Studies of Muir Glacier, Alaska, Nat. Geog. Mag., Vol. IV, 1892, pp. 19–84; Glacier Bay and Its Glaciers, 16th Ann. Rept., U. S. Geol. Survey, Part I, 1896, pp. 421–461.

[10] Amer. Geol., Vol. VIII, 1891, pp. 207–230.

[11] Amer. Geol., Vol. IX, 1892, pp. 190–197.

[12] Atlas of Award, Alaskan Boundary Tribunal, Sheets 14, 15, 16, 17, 18, 19.

[13] See H. F. Reid's Variations of Glaciers, Jour. Geol., Vol. V, 1897, p. 381.

The Harriman Alaska Expedition visited Glacier Bay in 1899 and John Muir and G. K. Gilbert studied the glaciers and glaciation.[1] Andrews and Case [2] made brief observations of Muir Glacier in 1902, and a United States Geological Survey Expedition in 1906, under charge of F. E. and C. W. Wright, made detailed studies and maps, as yet only published in abstract.[3] In 1907 the Boundary Survey work resulted in additional observations which have been briefly described by Klotz [4] and by Morse [5] and more recently published in a new United States Coast and Geodetic Survey chart [6] whose comparison with that of 1892 [7] shows the great glacial retreat that has been going on in this region and which has been summarized by H. F. Reid.[8] In 1911 Glacier Bay was visited by the authors of this book [9] who found most of the ice tongues retreating, though Rendu Glacier had advanced 1½ miles. In 1912 the Grand Pacific Glacier was mapped by N. J. Ogilvie of the Canadian Boundary Survey, who found that it retreated 1¼ miles during two months, making a new Canadian harbor.[10] In 1913 Glacier Bay was visited by an excursion of the International Geological Congress.[11] Afterwards the junior author of this book studied several of the ice tongues of Glacier Bay, finding that Grand Pacific Glacier had readvanced and destroyed the new Canadian harbor,[12] and that the Reid, Lamplugh, and de Margerie Glaciers had also advanced. A great number of popular descriptions of Muir and other Glacier Bay ice tongues have been written.[13]

Davidson has summarized the information based upon the maps and reports of Vancouver, Tebenkof, and the Russian navigators and agrees with Wright and Reid that Muir Glacier and the other ice tongues of Glacier Bay extended 25 to 40 miles farther to the south in 1794. He also refers to native legends of the greater extension of these glaciers.[14]

Muir Glacier receded about 1½ miles between Muir's observations of 1879–1880 and Reid's in 1890 and the several fronts of Grand Pacific Glacier retreated from one to six miles. Between 1890 and 1892 Muir Glacier advanced about 300 yards, retreating about the same amount between 1892 and 1894. Between 1879 and 1896 Rendu and Carroll Glaciers retreated 2 to 4 miles. Between 1899 and 1907 Muir Glacier retreated about 8½ miles and Grand Pacific Glacier about 8 miles. The rapid retreat of the

[1] Harriman Alaska Expedition, Vol. III, 1904, pp. 16–39.

[2] Nat. Geog. Mag., Vol. XIV, 1903, pp. 441–444.

[3] Journ. Geol., Vol. XVI, 1908, pp. 52–53.

[4] Klotz, Otto, Geog. Journ., Vol. XXX, 1907, pp. 419–421.

[5] Morse, Fremont, Nat. Geog. Mag., Vol. XIX, 1908, pp. 76–78.

[6] Chart 8306.

[7] Chart 3095.

[8] Variations of Glaciers, Journ. Geol., Vol. IX, 1901, p. 253; X, 1902, p. 317; XI, 1903, p. 287; XII, 1904, pp. 258–260.

[9] See H. F. Reid's Variations of Glaciers, Journ. Geol., Vol. XXI, 1913, pp. 425–426.

[10] Martin, Lawrence, Glaciers and International Boundaries, Scientific American Supplement, Vol. LXXVI, 1913, pp. 129, 136–138.

[11] Martin, Lawrence, Guide Book No. 10, Excursion C 8, International Geological Congress, Ottawa, 1913, pp. 121–132, and table of errata.

[12] Martin, Lawrence, The Literary Digest, Vol. XLVII, No. 1229, Nov. 8, 1913, p. 871, and three maps.

[13] Burroughs, John, Harriman Alaska Expedition, Vol. I, 1902, pp. 35–48.

Egerton, H. G., Alaska and Its Glaciers, Nineteenth Century, Vol. 22, pp. 997–999.

[14] Davidson, George, Trans. and Proc. Geog. Soc. Pacific, Vol. III, 1904, pp. 66–72.

recent period began just after the Harriman Expedition's observations in June, 1899, and seems to have been initiated, and perhaps somewhat hastened, by the shaking during the severe Yakutat Bay earthquakes of September, 1899, but in large part as a result of the increased area of ice front exposed to iceberg discharge. The rapid retreat of Muir Glacier has resulted in the loss of much of its scenic beauty and in its relative inaccessibility to tourist steamers between 1899 and 1909 because of the increased amount of floating ice. This retreat is also dismembering the glacier systems so that the number of separate ice fronts is continually increasing.

Alexander Archipelago. Chicagof, Baranof, Kruzof, Admiralty, Kupreanof, and Prince of Wales Islands and the smaller islands of the Alexander Archipelago seem to have all been glaciated but only small glaciers are known to linger upon them. Brooks shows two small glaciers upon Baranof Island east of Sitka [1] on one of his maps and Knopf has alluded to several small glaciers here.[2]

Glaciers on Pacific Coast of Fairweather Range.[3] There are a number of ice tongues that flow south or west from the Fairweather Range either to Cross Sound and Icy Strait or to the Pacific. Among these is the Brady Glacier, a through glacier heading on the divide with Reid Glacier of Glacier Bay and having a length of 27 miles and a width of 2½ to 6 miles. It is not tidal but ends within 2 miles of Taylor Bay. It advanced 5 miles between 1794 and 1894,[4] and Muir reports that it was advancing and destroying trees in 1880.[5]

LaPerouse Glacier is a small piedmont glacier near the base of Mt. Fairweather, and and the only Alaskan ice tongue that enters the open ocean and discharges icebergs at present. It was visited and studied carefully by G. K. Gilbert in June, 1899,[6] when he found that it had retreated one or two hundred yards from a forest edge into which it was advancing in September, 1895, as shown by a photograph from a Fish Commission vessel. The junior author on October 5, 1904, and both of us in September, 1909, saw LaPerouse Glacier from a steamer close in shore and it had then retreated still farther from the forest than when Gilbert was there ten years before. The same was true in 1906 when the Wright brothers of the United States Geological Survey were there.[7] Between September 4, 1909, and June 10, 1910, when the junior author was fortunate enough to see the glacier from another vessel close inshore, the LaPerouse Glacier had advanced across the barren zone of three to six hundred feet observed by Gilbert, plus the thousand feet or so of additional retreat during the ensuing ten years, and was once more destroying the forest on both margins, as it had been fifteen years before.[8] The glacier next east of LaPerouse had also retreated from a forest edge in 1899 but was not advancing in 1909 or 1910.

[1] Brooks, A. H., Professional Paper 45, U. S. Geol. Survey, 1906, Pl. XII.

[2] Bull. 504, U. S. Geol. Survey, 1912, pp. 10–11.

[3] U. S. Coast and Geod. Survey, Charts 3089, 8050, and 8304; Coast Pilot, Alaska, Part I, 1908, Pl. opposite p. 184.

Atlas of Award, Alaskan Boundary Tribunal, Sheets 15, 16, 19.

[4] Klotz, Otto, Geog. Journ., Vol. XIV, 1899, pp. 526–528.

[5] Reid, H. F., Variations of Glaciers, Journ. Geol., Vol. V, 1897, p. 380.

[6] Harriman Alaska Expedition, Vol. III, 1904, pp. 39–45.

[7] Personal Communication from F. E. Wright.

[8] Martin, Lawrence, Gletscheruntersuchungen längs der Küste von Alaska, Petermanns Geog. Mitt., Jahrgang 1912, Augustheft, p. 78; Journ. Geol., Vol. XIX, 1911, p. 457.

The tidal glaciers that enter Lituya Bay have advanced 2½ to 3 miles between 1786, when mapped by LaPerouse, and 1894 when mapped by the Canadian Boundary Survey.[1] Dall[2] thinks at least a mile of this advance was between 1874 and 1894. One of these glaciers advanced another half mile between 1894 and 1906.[3]

Farther west is the Grand Plateau Glacier, formerly thought to be nearly as large as the Malaspina and Bering Glaciers. It is quite unexplored but is plainly visible from the steamer. It was named by LaPerouse and described by Vancouver[4] and by Dall.[5] What was erroneously mapped by the Boundary Survey in 1895 as one large piedmont glacier was clearly seen by the junior author in 1913 to be at least two separate ice masses.

Glaciers of the Lower Alsek River. These ice tongues include at least one through glacier heading eastward on the Glacier Bay divide and a number of through glaciers heading westward on the divide with the glaciers of Russell Fiord. They were first seen by Glave who came down the Alsek River in 1890.[6] He described five great glaciers, of which the lowermost, Alsek Glacier, is the only one bearing a name. One of these is the northern terminus of the through glacier called Grand Pacific Glacier in Glacier Bay. At least one other is a through glacier whose western termini are Nunatak and Hidden Glaciers of Russell Fiord and Yakutat Glacier, east of Yakutat Bay.

The Alsek Glacier and part of the glacier system west of the river were mapped by the Boundary Survey in 1895.[7] In 1906 Blackwelder and Maddren visited, mapped, and briefly described the Alsek Glacier as well as the Yakutat Glacier and smaller ice tongues of the mountain front between Alsek River and Russell Fiord,[8] several of which had been roughly mapped by a United States Fish Commission party in 1901.[9] In 1906 and 1908 the Boundary Commission parties ascended the Alsek, making more detailed maps and observations of the ice tongues surveyed by Glave and the Canadian Boundary Survey party of 1895. They found[10] the northern terminus of Grand Pacific through glacier retreating while the glacier at the upper canyon was advancing and crushing alder bushes. The glacier next above Alsek Glacier, a terminus of the Nunatak-Hidden-Yakutat through glacier system, was 2½ miles back from the river and thoroughly stagnant. Alsek Glacier was more active in 1908 than in 1906.

The ice-flooded region between Alsek River and Russell Fiord is a typical glaciated region, with all the valleys filled with ice and many detached ridges and peaks rising above the glacier and snowfield.[11] It is a distance of 39 to 42 miles by the through glacier sys-

[1] Klotz, Otto, Geog. Journ., Vol. XIV, 1899, pp. 524–526.

[2] Dall, W. H., in H. F. Reid's Variations of Glaciers, Journ. Geol., Vol. VII, 1899, p. 225.

[3] Wright, F. E. and C. W., Journ. Geol., Vol. XVI, 1908, p. 53.

[4] Vancouver, Capt. George, Voyage of Discovery, Vol. V, London, 1801, pp. 358–359.

[5] Dall, W. H., Journ. Amer. Geog. Soc., Vol. XXVIII, 1896, p. 3.

[6] Glave, E. J., Leslie's Illustrated Newspaper, Vol. 71, 1891; Century Magazine, Vol. XXII, 1892, p. 880.

[7] Atlas of Award, Alaskan Boundary Tribunal, Sheet 20; Pl. VII, Professional Paper 64, U. S. Geol. Survey, 1909.

[8] Blackwelder, Eliot, Journ. Geol., Vol. XV, 1907, pp. 415–433.

[9] Pl. XLIII, Bull. 21, U. S. Fish Commission, 1902.

[10] Morse, Fremont, and Netland, L., Journ. Geol., Vol. XVII, 1909, pp. 669–670; see also photographs, Nat. Geog. Mag., Vol. XX, 1909, pp. 597, 598, 605.

[11] Atlas of Award, Alaskan Boundary Tribunal, Sheets 20 and 21.

tem from the unnamed second glacier of Alsek River to the terminus of Hidden or Nunatak Glacier of Russell Fiord, the valley glaciers averaging two to five miles in width and reaching elevations of from 2000 to 3000 feet on the highest divides.[1]

This region is one of glacier highways from Yakutat Bay to the Alsek over the Nunatak or Fourth Glaciers. Several hundred prospectors utilized the Nunatak (Third Glacier) route in 1898, among them the Messrs. Hill from whose sketches the United States Geological Survey has compiled a map,[2] which shows the seven or more great valley glaciers reaching the western or Kaskawulsh branch of Alsek River, and the otherwise unexplored glaciers east of Mounts Seattle and Hubbard. A complex system of through glaciers connects the Upper Alsek glaciers with the Nunatak and Hubbard Glaciers of the Yakutat Bay region, the distance over the ice being forty miles or more.

Glaciers of Yakutat Bay and Mt. St. Elias. The Hidden, Fourth, Nunatak, Variegated, Hubbard, Haenke, Turner, Black, Galiano, Atrevida, and Lucia Glaciers of Yakutat Bay, Disenchantment Bay, and Russell Fiord, and the Hayden, Marvine, Seward, Agassiz, Libbey, Tyndall, and Guyot tributaries of Malaspina Glacier are to be so fully described in later chapters that they will not be further discussed here.

Glaciers North of the St. Elias Range. Throughout this loftiest part of the St. Elias Range, 10,000 to 19,000 feet, whose glaciers are still largely unexplored, there is an intricate system of through glaciers filling the valleys and passes. For 140 miles from Alsek to Copper River no one has ever yet crossed the range. A striking feature is the smaller number and size of glaciers on the north or leeward side of the range in contrast with the great number of large glaciers that cover the south or windward side in the Yakutat Bay and Mt. St. Elias regions.

These ice-filled valleys of the main St. Elias Range include unexplored through glacier systems that extend from Yakutat Bay to the upper Alsek as well as to the headwater regions of the White, the Tanana, and the Copper Rivers. The Abruzzi party in 1897 named the Colombo, A. Q. Sella, and other glaciers north of Mt. St. Elias.[3] The Kaskawulsh or O'Connor or Slims River Glacier is perhaps the northern terminus of the Hubbard through glacier system.[4] Its two streams flow one to the Alsek and directly to the Pacific Ocean, the other to White and Yukon Rivers and Bering Sea. In 1899 and 1904 Kaskawulsh Glacier was retreating.[5] Other north-flowing ice tongues of the St. Elias Range are the two Donjek River glaciers, 20 to 23 miles long, the Klutlan Glacier, and the Russell Glacier, shown on several maps[6] and briefly described by Hayes in 1891,[7]

[1] Unpublished Map, 1906, Sheet 2846, Archives U. S. Coast and Geodetic Survey.

[2] Alsek River Region, 1: 360,000, 1905, unpublished.

[3] The Ascent of Mt. St. Elias, London, 1900, pp. 156–160.

[4] Map of Kluane, White, and Alsek Rivers, Yukon Terr., Dept. of Int., Canada, 1905.

[5] Brooks, A. H., 21st Ann. Rept., U. S. Geol. Survey, Part II, 1899–1900, Pl. XLIII, Pl. XLV A, and p. 364. McConnell, R. G., the Kluane Mining District, Geol. Survey of Canada, Ann. Rept., 1904, Vol. XVI, pp. 2A, 9A, 10A, and Map 894.

[6] Nat. Geog. Mag., Vol. IV, 1892, Pl. 20.
21st Ann. Rept., U. S. Geol. Survey, 1899–1900, Part II, Pl. XLIII.
Map of Kluane, White and Alsek Rivers, Yukon Terr., Dept. of Int., Canada, 1905.
Professional Paper 41, U. S. Geol. Survey, Pl. XX.
Chitina Quadrangle, Map 601, Alaska, U. S. Geol. Survey; Nizina Special Quadrangle, Map 601B, Alaska, U. S. Geol. Survey.

[7] Hayes, C. W., Nat. Geog. Mag., Vol. IV, 1892, pp. 151–154.

by Brooks in 1899 [1] and by Moffit and Capps in 1908,[2] during which time they have apparently been retreating uniformly. Logan Glacier, northwest of Mt. St. Elias at the headwaters of the Chitina River, was discovered by the Boundary Survey in 1912 when it was advancing.

Glaciers of the Wrangell Mountains. One of the most compact systems of Alaskan glaciers is that upon the Wrangell Mountains, 10,000 to 16,000 feet, where the Drop, West, Copper, Jacksina, Nabesna, Chisana, and minor glaciers flow northward, while the Frederika, Nizina, Rohn, Regal, Kennicott, Kuskulana, Kluvesna, Long, Cheshnina, Chetaslina, Dadina, Nadina, and others flow south. Nabesna Glacier is 55 miles long and 2½ miles wide. Kennicott Glacier is 19 miles long and 2 to 3 miles wide. These glaciers have been partly or wholly described and mapped by Hayes in 1891, Rohn in 1899, Schrader and Spencer in 1900, Mendenhall and Schrader in 1902, Moffit, Maddren, and Capps in 1907 and 1908, and Dunn in 1908,[3] the latter being the first to ascend the 14,000 foot, snow-sheathed active cone of Mt. Wrangell. The snowfields of these glaciers nearly cover the Wrangell Mountains, which include four volcanoes, Mts. Wrangell, Drum, Sanford, and Blackburn. The lower portion of Kennicott Glacier and several adjacent ice tongues were accurately mapped in 1908 by D. C. Witherspoon and described by F. H. Moffit and S. R. Capps.[4] In 1911 and 1912 several of these glaciers were traversed by Miss Dora Keen in her ascent of Mt. Blackburn.[5]

The radial consequent glaciers of Mt. Sanford are notable, as is the unsymmetrical glacier drainage development on Mt. Wrangell, where it has been supposed that the greater heat on the steeper west side of this active volcano results in melting away the western glaciers. Some of the recent lavas overlie glacial deposits but have been themselves glaciated since cooling. Nabesna and Chisana Glaciers, northeast of Mt. Wrangell, were retreating in 1898, but Nizina and Russell Glaciers advanced between 1899 and 1908. Frederika Glacier was advancing when seen by Hayes in 1891 but had begun to recede before the visit of Moffit and Capps in 1908, while the small glacier opposite it was then advancing.

[1] Brooks, A. H., 21st Ann. Rept., U. S. Geol. Survey, 1899–1900, Part II, p. 364 and Pl. I.

[2] Capps, S. R., Journ. Geol., Vol. XVIII, 1910, pp. 48–55.

[3] Hayes, C. W., Nat. Geog. Mag., Vol. IV, 1892, pp. 153–154, and map, Pl. 20.

Schrader, F. C., 20th Ann. Rept., U. S. Geol. Survey, 1900, p. 377.

Rohn, Oscar, Copper River Exploring Expedition, Narratives of Explorations in Alaska, Washington, 1900, pp. 791, 799–801, 820; 21st Ann. Rept., U. S. Geol. Survey, Part II, 1900, pp. 406–407, and map, Pl. LII.

Schrader, F. C. and Spencer, A. C., House Doc. 546, 56th Congress, 2d Session, Washington, 1900, pp. 31–32, 58–61, 76–81 and map, Pl. II.

Mendenhall, W. C. and Schrader, F. C., Professional Paper 15, U. S. Geol. Survey, 1903, Pl. III.

Mendenhall, W. C., Professional Paper 41, U. S. Geol. Survey, 1905, pp. 18, 62–72, 88–90, and map, Pls. XIX and XX.

Mendenhall, W. C., Nat. Geog. Mag., Vol. XIV, 1903, pp. 395–407.

Moffit, F. H. and Maddren, A. G., Bull. 374, U. S. Geol. Survey, 1909, pp. 37–42, and map, Pls. I, II.

Dunn, Robert, Harper's Magazine, Vol. XCVIII, 1909, pp. 479–509.

Capps, S. R., Journ. Geol., Vol. XVIII, 1910, pp. 33–57; Science, N. S., Vol. 30, 1909, p. 974; Journ. Geol., Vol. XVIII, 1910, pp. 359–375; Bull. 417, U. S. Geol. Survey, 1910, pp. 36–42.

See also photographs, Nat. Geog. Mag., Vol. XX, 1909, pp. 612, 617, 618, 620, 621.

[4] Bull. 448, U. S. Geol. Survey, 1911, Pl. II and pp. 43–59.

[5] Keen, Dora, Scribner's Magazine, Vol. 52, 1912, pp. 64–80; Ladies' Home Journal, Vol. 30, August, 1913, p. 7.

Glaciers of the Eastern Chugach Mountains. The Chugach Mountains are a part of the St. Elias Range, forming the Coast Range westward from Malaspina Glacier to Cook Inlet, and are from 6000 to 10,000 feet in altitude. The Bering Glacier, a large piedmont glacier just west of Mt. St. Elias, and largely unexplored, is fed by unnamed tributaries from the south side of the Chugach Mountains and descends nearly to sea level, being fringed by a border of outwash fans from Yakataga to Controller Bay. Its area is probably between 1000 and 1500 square miles.

Sir Edward Belcher described some of the features of Bering Glacier in 1837,[1] and Seton Karr observed it in 1886.[2] Its general position has been shown upon several maps [3] and its western portion was carefully mapped [4] and described in 1903–1906 by Martin [5] who also briefly described the adjacent Martin River Glacier, the Slope Glacier and the small glaciers of Ragged Mountain near Katalla.

Glaciers of the Lower Copper River. The Miles, Childs, Sheridan, and adjacent glaciers of the front of Chugach Mountains and lower Copper River are not discussed here because described fully in later chapters.

Farther up Copper River the glaciers on the north side of the eastern Chugach Mountains are largely unexplored. One of them is Tana Glacier of the Nizina headwaters, a small north or leeward-slope glacier, which forms a striking contrast to the great Bering Glacier of the south slope. Schwan and Woodworth Glaciers [6] drain into the Copper by Tasnuna River[7] and there are smaller ice tongues as at Cleave Creek, Worthington Glacier, east of Valdez, the ice tongues at the head of Bremner River, and many others.

Glaciers of Prince William Sound. The Valdez, Shoup, and Columbia Glaciers, and the ice tongues of College Fiord, Harriman Fiord and other portions of Prince William Sound are to be fully described in later chapters.

Glaciers North of the Chugach Mountains. As in the case of the main St. Elias Range this part of the same range shows fewer ice tongues on the north or leeward side than those that descend on the ocean side to Prince William Sound. Many of them are of good size, however, as the Knik, Matanuska, Tazlina, and others. These glaciers were first mapped and described by Mendenhall and Glenn in 1898, by Martin in 1905 and by Paige an d Knopf in 1906.[8] Several of the larger ones may be through glaciers connecting with

[1] Quoted by Blake, Amer. Journ. Sci., 2d series, Vol. XLIII, 1867, pp. 100–101, and by Davidson, Trans. and Proc. Geog. Soc., Pacific, Vol. III, 1904, pp. 42–43.

[2] Seton Karr, H. W., Proc. Roy. Geog. Soc., Vol. IX, 1887, pp. 269, 274, 276; Shores and Alps of Alaska, London, 1887, pp. 138–142.

[3] U. S. Coast and Geod. Survey, Charts 8502 and 8513.
U. S. Geol. Survey, Bull. 335, 1908, Pl. I.

[4] Map of Controller Bay Region, No. 601 A, U. S. Geol. Survey, also in Bull. 335, Pl. II.
Chitina Quadrangle, Map 601, U. S. Geol. Survey; also in Bull. 374, Pl. I.

[5] Martin, G. C., Bull. 250, U. S. Geol. Survey, 1905, pp. 16–17; Bull. 335, 1908, pp. 16, 46–54, 64–65, and map, Pls. II and V.

[6] Schrader, F. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1899, pp. 362, 364, 396–398 and map, Pls. 20 and 21.

[7] Chitina Quadrangle, Map 601, U. S. Geol. Survey; also in Bull. 374, Pl. I.

[8] Mendenhall, W. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 326–327 and map, No. 16.
Glenn, E. F., War Dept , Adj. Gen. Office, No. XXV, 1899, p. 58 and map in pocket.
Martin, G. C., Bull. 289, U. S. Geol. Survey, 1906, pp. 7, 15, and map, Pl. I.
Paige, S. and Knopf, A., Bull. 327, U. S. Geol. Survey, 1907, pp. 33–37 and map, Pl. I.

the ice tongues of Prince William Sound, Tazlina Glacier perhaps connecting with Columbia Glacier, Matanuska Glacier with Harvard and Yale Glaciers, and Knik Glacier with Barry or some other glacier of Harriman Fiord.

Just north of the Chugach Mountains are the Talkeetna Mountains, 5000 to 7500 feet high, within which are a series of smaller glaciers, the Chickaloon, Talkeetna, and others, observed by soldiers accompanying Captain Glenn,[1] by Martin,[2] and by Paige and Knopf, and shown on the same maps as those of the northern Chugach.

Glaciers of the Kenai Peninsula. The Kenai Mountains, which continue the Chugach Mountains to the southwest and form the west end of the St. Elias Range, rise to heights of 6000 to 7000 feet and send valley glaciers east to Prince William Sound, south to the Gulf of Alaska, and west toward Cook Inlet. There are many tidewater glaciers in western Prince William Sound and on the south of the Kenai Mountains, but none that discharge directly into Cook Inlet on the west.

The ice tongues of western Prince William Sound include the Portage Glacier, the glaciers of Blackstone Bay, Port Nellie Juan, and Icy Bay. There are glaciers on the south side of Kenai Peninsula in Port Bainbridge, Fairfield and Days Harbors, Resurrection Bay, Aialik and Nuka Bays, and Port Dick. In the interior of the peninsula small glaciers discharge into Glacier River,[3] into the stream feeding Lake Skilak, and into Lake Tustumena. In the southwestern part of the peninsula the Grewingk, Wosnessenski, Doroshin, and Southern Glaciers descend nearly to Kachemak Bay and Cook Inlet. Several of these are through glaciers, as the Portage Glacier of Passage Canal and Turnagain Arm, the Southern Glacier of Tutka Bay and Port Dick, etc.

A number of the glaciers of the south coast of Kenai Peninsula were discovered by the Russians and are shown on Tebenkof's chart of 1849. Those in western Prince William Sound were mapped by Vancouver's lieutenant, Whidbey, in 1794 [4] and by Applegate in 1887.[5] The glaciers of Kachemak Bay were visited by Dall in 1880, by Dall,[6] Becker and Curtiss [7] in 1895, and by Gilbert and the Harriman Expedition in 1899.[8] Grewingk Glacier retreated about 250 feet between 1880 and 1895, and 350 feet between 1895 and 1899.

Dall and Becker [9] have alluded to the evidences of former glaciation in Kodiak and Wood Islands, the westward extension of the Kenai Mountains, as amplified later by Gilbert.[10] No glaciers are known on Kodiak Island.

[1] Mathys, F. and Bagg, J. S., Compilation of Narratives of Exploration in Alaska, Washington, 1900, pp. 681, 682–683.

[2] Martin, G. C. and Katz, F. J., Bull. 500, U. S. Geol. Survey, 1912, pp. 25–26, 67–72.

[3] Tarr, R. S. and Martin, Lawrence, An Effort to Control a Glacial Stream, Annals Assoc. Amer. Geog., Vol. II, 1912, pp. 25–40.

[4] Vancouver, Capt. George, Voyage of Discovery, London, Vol. V, 1801, pp. 313–314.

[5] Davidson, George, Trans. and Proc. Geog. Soc., Pacific, Vol. III, 1904, pp. 18–20, 22–28, and Maps IV and V.

[6] Dall, W. H., Journ. Amer. Geog. Soc., Vol. XXVIII, 1896, p. 15; 17th Ann. Rept., U. S. Geol. Survey, Part I, 1896, Pl. LI; 18th Ann. Rept., U. S. Geol. Survey, Part III, 1898, Pl. XIII; Coast and Geod. Survey, Chart 8651; Variations of Glaciers, Journ. Geol., Vol. V, 1897, p. 381.

[7] Curtiss, F. H., in H. F. Reid's Variations of Glaciers, Journ. Geol., Vol. VI, 1898, pp. 475–476.

[8] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 97–102.

[9] Becker, G. F., 18th Ann. Rept., U. S. Geol. Survey, Part III, 1898, pp. 59–60.

[10] Harriman Alaska Expedition, Vol. III, 1904, pp. 177–182.

Some of the glaciers of the interior of Kenai Peninsula [1] were observed by Mendenhall in 1898 and Moffit in 1904.[1] In 1908 and 1909 Grant mapped the fronts of many of the glaciers of eastern and southern Kenai Peninsula.[2] The former were studied in 1910 by the junior author of this book, and are described in later chapters.

The Alaska Range

Glaciers of the Nutzotin and Mentasta Mountains. The eastern end of the Alaska Range includes the Nutzotin and Mentasta Mountains which rise to heights of 5000 to 10,000 feet. Being relatively low and in a region of low precipitation, because in the lee of the higher Chugach and Wrangell Mountains of the St. Elias Range, they nourish only a few small glaciers, some of which were seen by Brooks in 1899,[3] by Schrader and Witherspoon in 1902 [4] and by Moffit, Knopf, and Capps in 1908.[5]

Glaciers of the Copper River Headwaters. The region between Mt. Kimball and Mt. Hayes in the Alaska Range rises to heights of from 8000 to 13,000 feet. It must receive more snowfall than the same range farther to the east, in the lee of Wrangell Mountains, for it has large snowfields and sends good-sized glaciers southward into the Copper River valley and smaller ones northward into the drainage basin of the Tanana and Yukon. These ice tongues include the Chistochina, West Fork, Gakona, Gulkana, Canwell, and Castner Glaciers. The first three glaciers drain to Copper River and the Pacific Ocean, the last two to the Tanana and Yukon Rivers and Bering Sea, while Gulkana Glacier sends one stream each way.

Gakona Glacier is a simple, single, consequent, valley glacier 13 miles long and 3 miles wide and with its direction of flow at right angles to the axis of the Alaska Range. West Fork Glacier is similarly simple. East of Mt. Kimball is a series of three small consequent valley glaciers, each about five miles long, which enter an east-west valley parallel to the axis of the range. The easternmost of the three glaciers turns eastward in an unsymmetrical piedmont bulb a mile long, draining to Slana River. The westernmost similarly turns westward in a one-sided piedmont bulb a mile long, draining to the Middle Fork of Chistochina River. The middle glacier makes a symmetrical bulb whose drainage also goes westward under the last-named glacier. These are consequent valley glaciers terminated by piedmont bulbs which divide a subsequent valley into two compartments.

West of these little glaciers is Chistochina Glacier, a piedmont through glacier in a later stage of development. It heads in two valleys, Middle and West Forks of Chistochina River. It is 10 miles long, a mile and a half to two miles wide and extends east and west paralleling the axis of the Alaska Range, from which it is fed by seven consequent valley glaciers two to four miles long, and by four smaller tributaries from the lower hills on the south. This seems to be a subsequent glacier.

[1] Mendenhall, W. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 328–329.
Moffit, F. H., Bull. 277, U. S. Geol. Survey, 1906, pp. 31–32 and map, Pl. II.

[2] Grant, U. S. and Higgins, D. F., Coastal Glaciers of Prince William Sound and Kenai Peninsula, Bull. 526, U. S. Geol. Survey, 1913, pp. 52–72, see Bull. 379, U. S. Geol. Survey, 1909, Pl. IV; Journ. Geol., Vol. XVII, 1909, pp. 670–671, and Bull. Amer. Geog. Soc., Vol. XLIII, 1911, pp. 721–737.

[3] Brooks, A. H., 21st Ann. Rept., U. S. Geol. Survey, Part II, 1901, p. 364.

[4] Professional Paper 41, U. S. Geol. Survey, 1905, Pl. XX.

[5] Moffit, F. H. and Knopf, A., Bull. 379, U. S. Geol. Survey, 1909, pp. 162, 169.
Capps, S. R., Journ. Geol., Vol. XVIII, 1910, p. 36.

The Canwell and Castner Glaciers were seen and mapped by Mendenhall and Glenn in 1898,[1] and briefly visited by the authors of this book in 1911. Those to the east were mapped and described by Mendenhall and Gerdine in 1902.[2] West of the Canwell and Castner Glaciers, which terminate in the valley of Delta River, are a number of large ice tongues near Mt. Hayes. They have been mapped by the U. S. Geological Survey. Those on the southern side of the Alaska Range have been briefly described by Moffit[3] and those on the northern side by Capps.[4]

Glaciers of the Mt. McKinley Region. This portion of the Alaska Range, between Mt. Hayes and Lake Clark, averages 7000 to 10,000 feet throughout its length and rises to 17,000 in Mt. Foraker and 20,300 feet in Mt. McKinley, the latter being the highest peak in North America. The east end, near Mt. Hayes, is largely unexplored; the west end has only small glaciers; but the central portion, near Mt. McKinley, supports some large valley glaciers. Those on the southeast slope are larger than those on the northwest. The ice tongues on the northwest include the Herron, Rearburn, Shainwald, Peters, Muldrow, and Harvey Glaciers. On the southeast side are the Fidele, Ruth, Kahiltna, Tokichitna, Caldwell, Fleishmann, Hayes, Stoney, and many unnamed glaciers. Several of these glaciers on the south are 30 or 40 miles long.

These glaciers have been observed and mapped by Dickey in 1897, Eldridge, Muldrow, and Spurr in 1898, Herron in 1899, Brooks in 1902, Cook and Dunn in 1903, Cook, Parker, and Porter in 1906.[5] All have been retreating so far as observed. Several were utilized as highways by Dunn and Parker in their unsuccessful attempts to climb Mt. McKinley in 1903 and 1906, as well as by Parker and Brown and by Stuck in their successful ascents of the mountain in 1912 and 1913.

Alaska Peninsula and Aleutian Islands

Glaciers of the Chigmit Mountains. These low mountains, 2000 to 5000 feet, have large glaciers practically only where volcanoes rise to a sufficient height to have snowfields. Of these active volcanoes Mts. Redoubt and Iliamna, 11,000 and 12,000 feet

[1] Mendenhall, W. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1901, pp. 327–328, 330–331, and Map 16; also photograph, Nat. Geog. Mag., Vol. XX, 1909, p. 616.

Glenn, E. F., War Dept., Adj. Gen. office, No. XXV, 1899, pp. 1–72 and map.

[2] Professional Paper 15, U. S. Geol. Survey, 1903, p. 58 and Pl. IX; Professional Paper 41, 1905, pp. 89–90, and map, Pl. XIX.

[3] Moffit, F. H., Bull. 498, U. S. Geol. Survey, 1912, pp. 39–45, 51–53, and Pl. I.

[4] Capps, S. R., Bull. 501, U. S. Geol. Survey, 1912, pp. 34–41, and Pl. I, Journ. Geol., Vol. XX, 1912, pp. 428–430, 432–3.

[5] Dickey, W. A., Nat. Geog. Mag., Vol. VIII, 1897, pp. 322–327.

Eldridge, G. H., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, p. 8, and Map No. 3.

Muldrow, R., Nat. Geog. Mag., Vol. XII, 1901, pp. 312–313.

Spurr, J. E., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 252–253, and Maps 6 and 12.

Herron, J. S., War Dept., Adj. Gen. Office, No. XXXI, 1901, pp. 30, 49, and map.

Brooks, A. H., Science, N. S., Vol. XVI, 1902, pp. 985–986; Nat. Geog. Mag., Vol. XIV, 1903, pp. 30–35; Smithsonian Rept. for 1903, pp. 407–425; Professional Paper 45, U. S. Geol. Survey, 1906, map, Pl. XI; Prof. Paper 70, U. S. Geol. Survey, 1911, pp. 125–128, 134–136.

Dunn, Robert, The Shameless Diary of an Explorer, New York, 1907, two maps.

Parker, H. C., Review of Reviews, Vol. XXXV, 1907, pp. 49–58.

Porter, R. W., Reconnaissance Map of the Region South of Mt. McKinley, including the Yentna Mining District, U. S. Geol. Survey, 1910.

Capps, S. R., Bull. 534, U. S. Geol. Survey, 1913, pp. 36–45 and Pl. II.

respectively, maintain good-sized glaciers. There are, however, minor glaciers in the mountains proper, including the several small ice tongues that drain to Lake Clark,[1] and doubtless others; and there may be many more, as the cirques in the mountains east of Iliamna Lake[2] testify.

A Coast Survey chart shows a large tidal glacier in Redoubt Bay[3] half way up the west side of Cook Inlet, but no description of it has been printed. West of Mt. Redoubt, which doubtless has other glaciers, there are five glaciers upon Mt. Iliamna, several of which were seen by Dall in 1895 [4] and also observed by the junior author in 1904. These include one a mile wide, at the head of Tuxedni or Snug Harbor, which was stagnant and moraine-covered in 1904, two on its east slopes, one a double glacier with a red moraine, the other with trees growing upon it, another near Chinitna Bay, and still another whose presence was known by its milky glacier stream; there are also several glacierlets on the St. Augustine volcano.

Glaciers of the Alaska Peninsula. Several glaciers descend from the east slope of Mt. Douglas, at least six of which were seen by Dall and Becker in 1895.[5] The Coast Survey chart [6] shows some of the glaciers upon the slopes of Mt. Douglas. In 1904 the junior author observed several of the glaciers in this part of the Alaska Peninsula. There were three large glaciers on the north slope of Mt. Douglas, 11,000 feet, the easternmost of which then ended a mile from sea level at an elevation of 150 feet. Six other large glaciers were seen upon the east and southeast slopes of Mt. Douglas, the large one just west of Cape Douglas having a bulb five or six miles wide and projecting 3 or 4 miles from the coast, although not discharging icebergs in 1904 because mantled with moraine at the terminus. Southwest of Cape Douglas is a large glacier with a remarkably sinuous medial moraine. The large glacier of Hello Bay was seen from a distance. All these glaciers seemed to be retreating in 1904. Pavlof Mountain has a large glacier seen in 1908 by Atwood,[7] who has also described the former glaciation of the adjacent islands and mainland.

On Katmai Pass, across the Aleutian Mountains of Alaska Peninsula, three or more small glaciers were seen by Spurr in 1900,[8] the larger ones being on the northwest side. South of Kialagvik Bay are three glaciers shown by Davidson [9] to be upon a map made by Vasilieff in 1831–1832 and seen by Dall in 1895 [10] and by the junior author from the steamer in September, 1904.

Glaciers of the Aleutian Islands. A few of the snow-capped volcanoes of the Aleutians are known to have radiating glaciers. There are ice tongues upon the slopes of Mts. Shishaldin, Round Top, and Isanof or Isanotski on Unimak Island, showing clearly in the photographs published by Westdahl.[11] The glacier of Mt. Makushin on the island of

[1] Osgood, W. H., Nat. Geog. Mag., Vol. XV, 1904, pp. 326–331.

[2] Martin, G. C. and Katz, F. J., Bull. 485, U. S. Geol. Survey, 1912, pp. 14–15, 82–94.

[3] Coast and Geod. Survey, Chart 8502, 1908.

[4] Dall, W. H., Journ. Amer. Geog. Soc., Vol. XXVIII, 1896, p. 11.

[5] Dall, W. H., Journ. Amer. Geog. Soc., Vol. XXVIII, 1896, pp. 8–9; 18th Ann. Rept., U. S. Geol. Survey, Part III, 1898, Pl. XIV.

[6] U. S. Coast and Geod. Survey, Chart 8502, 1908.

[7] Atwood, W. W., Bull. 467, U. S. Geol. Survey, 1911, pp. 13, 84–91.

[8] Spurr, J. E., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 254–255, and Map 11.

[9] Davidson, George, Trans. and Proc. Geog. Soc., Pacific, Vol. III, 1904, pp. 12–13, and Maps I and II.

[10] Dall, W. H., Journ. Amer. Geog. Soc., Vol. XXVIII, 1896, p. 17.

[11] Westdahl, F., Nat. Geog. Mag., Vol. XIV, 1903, pp. 91–99.

Unalaska was visited by Theodore Blake in 1867,[1] under the direction of Davidson.[2] When visited by Jagger in 1907 it showed no signs of retreat, nor did the glaciers on the island of Atka.[3] In 1889 Russell [4] visited Unalaska and concluded that there had been no general glaciation of the Aleutian Islands, although local glaciers had existed.

Central Plateau and Bering Sea

Yukon and Kuskokwim Valleys. The lack of former glaciation in the Central Plateau of Alaska was established by Dall,[5] Dawson,[6] McConnell,[7] Russell,[8] Hayes,[9] Spurr,[10] Tyrrell,[11] Brooks,[12] and others. Detailed topographic maps and geological investigations have failed to show existing glaciers within most of this great area, but that they were formerly present east of the Klondike district and in the upper Lewes, Teslin, Pelly and Macmillan Rivers is abundantly proved. In the region between the Tanana and Yukon Rivers the prospectors "Annual glaciers" referred to by Spurr,[13] Barnard [14] and Prindle [15] seem to be only long-lived snowbanks that melt before the summer is over. One region where these are abundant is called Glacier Mountain. Streams in the interior of Alaska which freeze solid and then overflow and freeze again are also known locally as "glaciers." The photographs from the Yukon-Tanana region show cirque-like amphitheatres at the heads of many streams and it seems likely that local glaciers formerly existed. This part of the Central Plateau rises to heights of 3000 to 6000 feet but has an annual precipitation of only 12 inches.

In part of the region between the Yukon and Koyukuk Rivers Mendenhall has shown [16] conclusively that there has been no glaciation but in another area between these rivers Eakin [17] found evidence of local glaciation in 1913.

In the Kuskokwim Mountains which rise to 4500 feet above sea level, there were formerly local glaciers, in a small area recently described by Maddren,[18] who finds cirques, moraines, and other conclusive evidence.

[1] House Extra Doc. 177, 40th Congress, 2d Session, 1868, p. 323.

[2] Davidson, George, Trans. and Proc. Geog. Soc., Pacific, Vol. III, 1904, pp. 9–11; Appalachia, Vol. IV, 1884.

[3] Journ. Geol., Vol. XVI, 1908, p. 668.

[4] Bull. Geol. Soc. Amer., Vol. I, 1890, pp. 138–180.

[5] Dall, W. H., Amer. Journ. Sci., Vol. XLV, 1868, p. 99.

[6] Dawson, G. M., Geol. and Nat. Hist. Survey Canada, Ann. Rept., Vol. III, Part I, 1889, p. 154B.

[7] McConnell, R. G., Bull. Geol. Soc. Amer., Vol. I, 1890, pp. 543–544.

[8] Russell, I. C., Surface Geology of Alaska, Bull. Geol. Soc. Amer., Vol. I, 1890, pp. 140–141.

[9] Hayes, C. W., Nat. Geog. Mag., Vol. IV, 1892, pp. 155–159 and Pls. 19 and 20.

[10] Spurr, J. E., 18th Ann. Rept., U. S. Geol. Survey, Part III, 1897, p. 270.

[11] Tyrrell, J. B., Bull. Geol. Soc. Amer., Vol. X, 1899, pp. 195–198.

[12] Brooks, A. H., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, p. 474; 21st Ann. Rept., U. S. Geol. Survey, Part II, 1901, pp. 364–365; Maps and Routes of Exploration in Alaska in 1898, Special Publication, U. S. Geol. Survey, 1900, pp. 69, 91.

[13] Spurr, J. E., 18th Ann. Rept., U. S. Geol. Survey, Part III, 1898, pp. 320–323.

[14] Barnard, E. C., Maps and Descriptions of Routes of Exploration in Alaska in 1898, Special Publication, U. S. Geol. Survey, 1899, p. 79.

[15] Prindle, L. M., Bull. 375, U. S. Geol. Survey, 1909, p. 12.

[16] Mendenhall, W. C., Professional Paper 10, U. S. Geol. Survey, 1902, pp. 45–46.

[17] Eakin, H. M., personal communication.

[18] Maddren, A. G., The Innoko Gold-Placer District, Alaska, Bull. 410, U. S. Geol. Survey, 1910, pp. 13–15, 58–60.

The Ahklun Mountains. The Ahklun or Oklune Mountains, a small group north of the Alaska peninsula near Bristol Bay are reported by Spurr [1] to still retain glacierlets, the one in cirque on Mt. Oratia being specifically mentioned. The mountains had previously been more fully occupied by local glaciers.

Islands of Bering Sea. In the shallow Bering Sea none of the islands, St. Lawrence, Nunivak, Hall, St. Matthew, and the Pribilof Islands, all of which are low and in a region of low precipitation, contain glaciers, and none seem to have ever been glaciated.[2]

THE ROCKY MOUNTAINS

Glaciers of the Endicott Mountains. The Rocky Mountains of northern Alaska, which reach an elevation of 5000 to 7000 feet in a latitude north of the Arctic circle, have an annual precipitation of only 6 to 12 inches and, therefore, maintain only small glaciers. One stagnant, dying glacier can be definitely referred to, an isolated débris-covered block observed in 1901 by Schrader and Peters [3] in the middle of the John River valley near the divide in latitude 68° north. It was then a detached circular mass 300 feet in diameter and 60 feet high. Schrader [4] speaks of other small glaciers seen by prospectors in the eastern part of the Endicott Mountains. Mendenhall reports the western Endicott Mountains as formerly glaciated,[5] and Maddren has described the present ice tongues and the former glaciation of the eastern Endicott Mountains, both in the part near the Hodzana Highland and Chandalar River [6] and in the region along the 141st meridian [7] between Porcupine River and the Arctic Ocean.

Baird Mountains. The Baird Mountains, which form the western extension of the Endicott Mountains, contain small glaciers, for they are from 2000 to 3000 feet high, higher than the mountains of Seward Peninsula, farther south, which still have small ice remnants. Smith [8] has described several of thesmall ice tongues and discussed the former glaciation. It seems quite certain that Baird Mountains were formerly extensively glaciated, for these mountain valleys must have fed the great valley glacier which extended down Kobuk River to the Bering Sea.[9]

Glaciers of the Seward Peninsula. Small glaciers are known to exist in the mountains of Seward Peninsula, which are 1800 to 2500 feet high, and receive 20 to 25 inches of precipitation yearly. The region is proved to have formerly had much more extensive

[1] Spurr, J. E., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 253–254 and Map No. 10.

[2] Dall, W. H., Alaska and its Resources, Boston, 1870, p. 462.

Muir, John, Report of the Cruise of the U. S. Revenue Steamer Corwin in the Arctic, 1881, pp. 133–145.

Dawson, G. M., Geological Notes on Some of the Coasts and Islands of Bering Sea and Vicinity; Bull. Geol. Soc. Amer., Vol. V, 1894, pp. 117–146.

Stanley-Brown, J., Geology of the Pribilof Islands: Bull. Geol. Soc. Amer., Vol. III, 1892, pp. 496–500.

Nordenskiold, A. E., The Voyage of the Vega, New York, 1882, pp. 583–585, 569.

Gilbert, G. K., Glaciers and Glaciation: Harriman Alaska Expedition, Vol. III, 1904, pp. 186–194.

[3] Schrader, F. C. and Peters, W. J., Professional Paper 20, U. S. Geol. Survey, 1904, pp. 84–91.

[4] Same, pp. 30 and 91.

[5] Mendenhall, W. C., Professional Paper 10, U. S. Geol. Survey, 1902, pp. 45–48.

[6] Maddren, A. G., Bull. 532, U. S. Geol. Survey, 1913, pp. 60–67.

[7] *Ibid*, personal communication.

[8] Smith, P. S., Bull Geol. Soc. Amer. Vol., 23, 1912, pp. 563–570.

[9] Hershey, O. H., Journ. Geol., Vol. XVII, 1909, pp. 83–91.

valley glaciers, as is testified by cirques, hanging valleys, transported bowlders, etc., in the Kigluaik, Bendeleben, York, Cape, and Ear Mountains. This is chiefly the result of studies by Brooks,[1] Collier, Smith, Eakin and Moffit.

The Glaciation of Alaska

The phenomena of former glaciation in Alaska have been treated incidentally in the preceding pages dealing with existing glaciers, and more fully in some of the reports there cited. Several reports, referred to below, have dealt especially with the general phenomena of Alaskan glaciation.

After Russell's visit to Alaska in 1889 he wrote his Notes on the Surface Geology of Alaska,[2] pointing out that northwestern North America had not been buried beneath a continental glacier when northeastern North America was ice-covered, but that in the appropriate part of Alaska "the ice age still lingers." At about the same time Dawson was writing his suggestive papers[3] on the glaciation of Yukon Territory and British Columbia in which he deals incidentally with parts of the Alaskan border, particularly from Mt. St. Elias to Dixon Entrance, and shows the extent of glaciation there.

After Gilbert's summer in Alaska with the Harriman Expedition in 1899, he published his wonderfully comprehensive descriptions of the existing glaciers observed, and then wrote on the Pleistocene glaciation, dealing especially with the physiographic forms produced by glaciation in the coastal region.[4]

In Brooks's Geography and Geology of Alaska[5] there is an excellent summary of the knowledge as to the glaciation of Alaska, accompanied by a map which showed the existing glaciers and snowfields, the glaciated areas, the névé fields and areas in part above ice sheets during maximum glaciation, the directions of ice movement, and the areas of post-glacial silts, sands, and gravels. Other writers have discussed glaciation,[6] and Hayes,

[1] Brooks, A. H., A Reconnaissance of the Cape Nome and Adjacent Gold Fields of Seward Peninsula, Alaska House Doc. 547, 56th Congress, 2d Session, Washington, 1901, pp. 43–47.

Brooks, A. H. and Collier, A. J., Science, N. S., Vol. XIII, 1901, pp. 188–189.

Collier, A. J., Professional Paper 2, U. S. Geol. Survey, 1902, pp. 28–29; Bull. 328, U. S. Geol. Survey, 1908, pp. 94–99.

Smith, P. S., Bull. 433, U. S. Geol. Survey, 1910, pp. 48, 135–137.

Smith, P. S. and Eakin, H. M., Bull. 449, U. S. Geol. Survey, 1911, pp. 83–86, 99.

Moffit, F. H., Bull, 533, U. S. Geol. Survey, 1913, pp. 14, 37, 61–63.

[2] Russell, I. C., Bull. Geol. Soc. Amer., Vol. I, 1890, pp. 99–154.

[3] Dawson, G. M., Recent Observations on the Glaciation of British Columbia and Adjacent Regions; Geog. Mag., 3d Decade, Vol. V, 1888, pp. 347–350; Amer. Geol., Vol. III, 1889, pp. 249–253; Geol. Nat. Hist. Survey Canada, Vol. III, Pl. 1, 1889, pp. 38B–43B; On the Later Physiographical Geology of the Rocky Mountain Region in Canada with Special Reference to Changes in Elevation and the History of the Glacial Period, Trans. Roy. Soc. Canada, Vol. VIII, sect. IV, 1890, pp. 25–74 and Pl. II, No. 4; Amer. Geol., Vol. VI, 1890, pp. 153–162.

[4] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, Glaciers, 1904, pp. 113–194; Nat. Geog. Mag., Vol. XV, 1904, pp. 449–450.

[5] Brooks, A. H., Professional Paper 45, U. S. Geol. Survey, 1906, pp. 245–249, and Pls. XII and XXII.

[6] Dall, W. H., Alaska and Its Resources, Boston, 1870, pp. 461–464.

Wright, G. F., Notes on the Glaciation of the Pacific Coast, Amer. Naturalist, Vol. XXI, 1887, pp. 250–256.

Tarr, R. S., The Glaciers and Glaciation of Alaska, Science, N. S., Vol. XXXV, 1912, pp. 241–258; Annals Assoc. Amer. Geog., Vol. II, 1912, pp. 3–24.

Spurr, and Brooks had partly shown its limits,[1] before the publication of Brooks's more complete 1906 map.

From this map it may be seen that the Pacific Mountains not only contain more and larger glaciers than the Arctic Rocky Mountains at present, but that the same was true at the stage of maximum glaciation. In the Yukon and Kuskokwim valleys and on the Arctic slope the glaciers may have extended short distances farther than the limits shown, for the till might be buried beneath outwash and recent alluvium, but the extension would be slight. On the Pacific Coast little is known of the seaward limits of the extended glaciers.

Extent and Importance of Alaskan Glaciers

The preceding brief outline of our knowledge of the present-day glaciers of Alaska and of the former extent of glaciers in that territory is probably not complete, but it suffices to show two facts with striking clearness,—first that there is an enormous number of glaciers in a very large area, and secondly that much remains to be done on the study of Alaskan glaciers and glaciation before we can claim to have more than the merest reconnaissance knowledge of the phenomena of this great field. When we consider the remoteness of the region, the inaccessibility of many parts of it, and the brief time in which systematic exploration has been in progress, the results already obtained in the study of the Alaskan glaciers and glacial phenomena are noteworthy, especially as the glacial study has often been but a minor incident in other work. To the United States Geological Survey, and especially to the Alaskan division, under the direction of Alfred H. Brooks, the major credit for these results is due.

From our outline of the present knowledge of Alaskan glaciers it is evident that it would be possible to give, even now, a much greater body of space to the subject, for in the reports to which reference has been made there are many important details and descriptions of individual glaciers. This has not been done, however, because even with the fullest abstracts of all reports, the discussion of Alaskan glaciers and glaciation would necessarily be most incomplete, with great and important gaps. It is our hope, in time, after some years of systematic work in the various glacier fields, to return to this subject and to present an outline description and discussion of living glaciers and glacial phenomena of Alaska, on the basis of wider observation. For scores of years, however, it is not to be expected that a complete account of the region will be possible. At present we do not know even the approximate number of Alaskan glaciers. There are several hundred to which names have been given; there are dozens which are tidal and other dozens which reach almost to the sea; scores reach beyond the mountain base and expand in piedmont bulbs; and scores and probably hundreds of glaciers have a length of from ten to fifty miles or more. In all, there are some thousands of glaciers in the Alaskan mountains only a small percentage of which have received names, and many of which have never even been seen.

The Alaskan region is one of the most wonderful regions of glaciation in the world, both from the standpoint of number and size of its glaciers, and from the extent and variety of

[1] Hayes, C. W., Nat. Geog. Mag., Vol. IV, 1892, Pls. XIX and XX.
Spurr, J. E., 18th Ann. Rept., U. S. Geol. Survey, Part III, 1897, Pl. XXXVII.
Brooks, A. H., 20th Ann. Rept., U. S. Geol. Survey, Part III, 1900, Map 25; 21st Ann. Rept., U. S. Geol. Survey, Part II, 1901, Pl. XLVII.

associated phenomena; and a thorough study of any of its facts is certain to yield important scientific results. The phenomena of advance and recession of the glacier termini, the former extent of the glaciers and their deposits, and the stupendous work which they have accomplished in sculpturing the wonderful series of fiords are among the phenomena demanding attention. The features exhibited have far more than local importance and application, for the fact that we have here large glaciers descending to sea level in a comparatively-warm, humid, north temperate climate gives rise to phenomena resembling those of the wasting margin of the great continental ice sheets of North America and Europe, and, therefore, throw light upon and furnish aid in interpreting these phenomena.

CHAPTER II

THE GLACIERS AND GLACIATION OF YAKUTAT BAY

General View of the Yakutat Bay Glaciers

Glacial Studies in Yakutat Bay. Among the many glaciers of Alaska those of a few areas have been singled out for more continuous and thorough study than others, either because of their accessibility or because they possess features of unusual interest. Of these areas three have attracted the greatest attention: (1) the glaciers of Glacier Bay and vicinity, (2) the glaciers of Yakutat Bay and vicinity, including the Malaspina Glacier; and (3) the glaciers of the Prince William Sound region, including those of lower Copper River. The National Geographic Society's Alaskan Expeditions of 1909 and 1910 studied the glaciers of the last two regions and this report deals mainly with the results of these studies, commencing, in this chapter, with the glaciers of Yakutat Bay and vicinity.

The results of the work in 1909–10, in so far as the Yakutat Bay region is concerned, are so intimately related to, and in large part dependent upon the results of previous studies that it will be necessary to briefly incorporate in this book the more significant points determined by the earlier studies, making this, therefore, a summary of several seasons' work.

The investigations of the glaciers of the Yakutat Bay region began in 1890, with Professor Russell's famous exploration of the Malaspina and adjacent glaciers in his first expedition[1] to Mt. St. Elias,[2] and was continued in 1891 in his second expedition to St. Elias.[3] Both before and after Russell's expeditions there were other explorations in this region, notably those of Schwatka, Topham, Bryant, and the Duke of the Abruzzi, but none that contributed much that bears directly upon the problems considered in this report,[4] until the Harriman Expedition in 1899.[5] The next visit of a scientific party having glacial study for its object was the United States Geological Survey Expedition of 1905 in charge of the senior author and to which the junior author was attached as special physiographic assistant.[6] The glaciers and glacial phenomena observed during this expe-

[1] Under the auspices of the National Geographic Society and the U. S. Geological Survey.

[2] Russell, I. C., An Expedition to Mt. St. Elias, Alaska; Nat. Geog. Mag., Vol. III, 1891, pp. 53–203.

[3] Russell, I. C., Second Expedition to Mt. St. Elias; Thirteenth Ann. Rept., U. S. Geol. Survey, pt. 2, 1892, pp. 1–91; Amer. Journ. Sci., Vol. XLIII, 1892, pp. 169–182; Journ. Geol., Vol. I, 1893, pp. 219–245.

[4] For bibliography of the Yakutat Bay region see I. C. Russell, Nat. Geog. Mag., Vol. III, 1891, pp. 58–74, and R. S. Tarr, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 20–25.

[5] Gilbert, G. K., Glaciers and Glaciation: Harriman Alaska Expedition, Vol. III, 1904, 231 pages; particularly pp. 45–70, 104, 170, and 209.

[6] A grant of money for this purpose was made by the American Geographical Society of New York.

dition are described in several publications.[1] In the reports of the work up to and including 1905 the glaciers of the Yakutat Bay region are fully described, and their more significant features interpreted. With a single exception, that of Galiano Glacier, the glaciers were found to be wasting, and recession was the rule between 1890 and 1905, as Russell's observations clearly show had been the general condition for years before his first visit of 1890.

Apparently having no bearing on the problems of the glaciers and glaciation, but in reality of the greatest significance, were the observations of 1905 upon the effects of the great earthquakes which disturbed the Yakutat Bay region in September, 1899.[2] For on returning to Yakutat Bay in 1906, the senior author, in charge of a second United States Geological Survey expedition, found that several of the hitherto-wasting glaciers had begun a sudden activity in response to an impulse derived from the down-shaking of great masses of snow in the glacier reservoirs during the earthquakes of 1899.[3]

The National Geographic Society's Alaskan Expeditions of 1909–10 went first to Yakutat Bay in order to determine what further changes had taken place in the interval since 1906. Since these later observations[4] fill out and verify those of the two previous expeditions, making, as it were, a complete story, it seems fitting to take up the Yakutat Bay region as a whole, describing and interpreting its glaciers and glacial phenomena in the light of all the studies from those of Russell in 1890 down to and including those of the season of 1913, when the junior author visited Yakutat Bay in charge of an excursion of the International Geological Congress.

Location and General Physiographic Features.[5] Yakutat Bay is an indentation in the otherwise almost unbroken coast line that stretches between Cross Sound and Controller Bay, its mouth being near latitude 59° 40′ north and longitude 140° west. It is about 40 miles southeast of Mt. St. Elias, and very near the point where the boundary line

[1] Tarr, R. S. and Martin, Lawrence, Observations on Glaciers and Glaciation of Yakutat Bay, Alaska: Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 99–101; Glaciers and Glaciation of Yakutat Bay, Alaska, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 145–167; Position of Hubbard Glacier Front in 1792 and 1794; Bull. Amer. Geog. Soc., Vol. XXXIX, 1907, pp. 129–136.

Tarr, R. S., Glacial Erosion in Alaska: Pop. Sci. Monthly, Vol. LXX, 1907, pp. 99–119.

[2] Tarr, R. S. and Martin, Lawrence: Recent Change of Level in Alaska: Science, Vol. XXII, 1905, pp. 879–880; Recent Changes of Level in the Yakutat Bay Region, Alaska: Bull. Geol. Soc. Amer., Vol. XVII, 1906, pp. 29–64; Recent Change of Level in Alaska; Geog. Journ., Vol. XXVIII, 1906, pp. 30–43; The Yakutat Bay Earthquakes of September, 1899, Professional Paper 69, U. S. Geol. Survey, 1912, pp. 11–30.

Martin, Lawrence, The Alaskan Earthquakes of 1899, Bull. Geol. Soc. Amer., Vol. XXI, 1910, pp. 339–407.

[3] Tarr, R. S., The advancing Malaspina Glacier: Science, Vol. XXV, 1907, pp. 34–37; Second Expedition to Yakutat Bay, Alaska; Bull. Geog. Soc. Phila., Vol. V, 1907, pp. 1–14; The Malaspina Glacier: Bull. Amer. Geog. Soc., Vol. XXXIX, 1907, pp. 273–285; Recent Advance of Glaciers in the Yakutat Bay Region, Alaska: Bull. Geol. Soc. Amer., Vol. 18, 1907, pp. 257–286. Also incorporated in the final report for the two expeditions of 1905 and 1906 in the section on Physiography and Glacial Geology of the Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 1–144.

[4] Tarr, R. S. and Martin, Lawrence, The National Geographic Society's Alaskan Expedition of 1909, Nat. Geog. Mag., Vol. XXI, 1910, pp. 1–54.

Martin, Lawrence, The National Geographic Society Researches in Alaska, Nat. Geog. Mag., Vol. XXII, 1911, pp. 537–561; The Hubbard Glacier, Pop. Sci. Monthly, Vol. 76, 1910, pp. 293–305; Gletscheruntersuchungen längs der Küste von Alaska, Petermanns Geog. Mitteilungen, Jahrgang 1912, pp. 78–79; Guide Book No. 10, Excursion C 8, International Geological Congress, Ottawa, 1913, pp. 134–162.

[5] For a more complete discussion of the physiography of the Yakutat Bay region see Tarr, R. S., Physiography and Glacial Geology, Professional Paper 64, U. S. Geol. Survey, 1909. For areal geology, see Tarr, R. S. and Butler, B. S., *ibid*, pp. 145–164.

between Canada and Alaska turns northward. The bay pierces first a low foreland occupied by the ice plateau of Malaspina Glacier on the west side, but on the southeast side composed of moraines and glacial gravels deposited during a former expansion of the Yakutat Bay and Russell Fiord glaciers. It then cuts across the low front range of mountains to the base of the lofty, snow-covered peaks of the St. Elias Range. It turns twice at high angles, and parallels itself back to the foreland, the distance from the mouth of the bay to its head being fully 75 miles.

The name Yakutat Bay is now applied only to the outer part of the inlet (Pl. I, A), mainly outside the mountain front, and throughout most of its extent bordered by a low coast, the wooded foreland on the southeast, the gravel fans and sandy beaches with strips of forest on the west, and behind these the moraine-covered ice bank which forms the outer border of Malaspina Glacier. This bay, which is 20 miles wide at the mouth, narrows progressively toward the head and, about half way between the mouth and head, the mountains approach to form the southeastern shore. At the head of the outer bay the mountains also converge from the west, and here the bay narrows to 3 miles. From this point, where the mountains form both shores of the inlet, on to the first bend the name Disenchantment Bay is applied. It is walled in by mountains rising abruptly from the sea to elevations of 2000 or 3000 feet and is a true fiord, with roughly parallel walls and a width of from 2½ to 4 miles. The name of the bay again changes where it bends abruptly at an acute angle, from here on to the head of the inlet being called Russell Fiord. Excepting at its very head this portion of the inlet is also a mountain-walled fiord with a width of from 1 to 3 miles. Where the northwestern arm of Russell Fiord bends slightly, at Cape Enchantment, the fiord bifurcates, one arm extending southward as the southeast arm of Russell Fiord, the other extending eastward as Nunatak Fiord.

With the exception of the irregular coast line of the foreland on the southeast shore of outer Yakutat Bay, and of the head of Russell Fiord, also in the foreland, the inlet, throughout almost its entire extent, has a generally straight coast line. There are a few small projecting points and a few small bays, but, although there are many valleys tributary to the inlet, the mouths of most of them are above the level of the fiord waters so that there are few bays. There are only two notable exceptions to this statement, Nunatak Fiord, at the head of which stands the ice cliff of Nunatak Glacier, and Seal Bay in the valley of which lies Hidden Glacier whose glacial streams have deposited an extensive outwash gravel plain, thereby greatly shortening this indentation. Were Hubbard Glacier to recede notably, there would doubtless be still other irregularities in the bay at the bend where Disenchantment Bay and Russell Fiord unite. The interpretation of these features, which are due to the erosive work of formerly expanded glaciers, may for the present be postponed.

The straight coast line which Yakutat Bay indents is a coastal plain built by glacial deposit, and mainly by glacial stream deposit, but worked over, in its outer portion at least, by the ocean waves which are ever beating upon its margin exposed to the open Pacific.[1] To the southeast of Yakutat Bay this coastal plain is still in process of active outbuilding, for numerous large glaciers descend from the mountains and spread out upon it at the mountain base, with huge glacial streams issuing from their fronts and flowing in braided courses across it; but these glaciers are evidently shrunken, and the

[1] Blackwelder, E., Amer. Journ. Sci., Vol. XXVII, 1909, pp. 459–466.

building of the plain is less active than formerly. To the northwest of Yakutat Bay the coastal plain is a mere fringe around the margin of the Malaspina Glacier and in two places is entirely absent, where the sea comes in contact with the glacier at Sitkagi Bluffs and Icy Bay. What the condition is beneath the Malaspina ice plateau is unknown.

In the immediate vicinity of Yakutat Bay the glaciers are greatly shrunken, and, excepting on the periphery of the Malaspina Glacier the construction of the foreland is progressing very slowly. Yet this is the broadest part of the coastal plain, testifying to the former great activity of deposit by glaciers and glacial streams. That the Yakutat Bay inlet exists in spite of the former activity of deposit is doubtless the result of the energetic erosion by the formerly expanded glaciers; and in the foreland area itself the existence of the bay is probably due to the occupation of the area by an expanded ice bulb which prohibited deposit there while it remained, and whose recession was so rapid that deposits failed to fill the depression. That deposit was in progress on the outer or ocean margin of the expanded bulb that filled Yakutat Bay is proved by the presence of a crescentic shoal completely across the mouth of the bay, on portions of which the ocean waves break in times of storm. Terminating, as it did, in the open ocean, the expanded Yakutat Bay Glacier was unable to deposit as rapidly here as in the case of glaciers ending on the foreland, for much of its supply of débris was borne off into the open ocean in icebergs and spread over a broader area than that reached by streams charged with glacial débris.

On the inner margin of the coastal plain the mountains rise abruptly. To the southeast of Yakutat Bay the Fairweather Range rises in magnificent proportions, almost from sea level, to elevations of from 10,000 to 15,000 feet or more, broken only by the glacier-skirt Alsek valley and by a series of broad glacier-filled troughs. With snowline descending to within 3000 or 4000 feet of sea level, with glaciers in all the valleys, and with its lofty peaks of varied form and elevation, this is one of the grandest mountain panoramas in the world with the whole mountain range, from crest to base, almost at sea level, exposed to view as one sails along the coast for a full day (see frontispiece). This panorama is perhaps not even rivalled by the St. Elias Range itself, which, though higher, lies farther away, with Malaspina Glacier between it and the sea. Between Mt. St. Elias and the Fairweather Range proper the lofty mountains of the continuous chain lie farther back from the coast and are separated from it by lower mountains and by the fringe of coastal plain already described. These foothills rise to no great height, rarely reaching elevations above 5000 feet; but their crests are snow-capped, while glaciers lie in most of the valleys; and through the larger ones descend great glaciers, fed among the higher mountains of the St. Elias Range farther inland. Both arms of the Yakutat Bay inlet pierce the foothills, making a complete cross section through them, and extending to the base of the crystalline mountains of the St. Elias Range. Here the mountains rise rapidly in elevation, reaching heights of from 10,000 to 16,000 feet within a few miles of the shore. From the shores of this bay the full grandeur of the St. Elias Range is seen perhaps at its best, and certainly from a central vantage point. Mt. Fairweather (15,330 feet) is visible to the southeast, the striking pyramid of St. Elias (18,024 feet) to the northwest, Mt. Logan (19,500 feet) to the north, Mt. Hubbard (16,400 feet) to the northeast, Mt. Vancouver (15,617 feet), Mt. Cook, (14,700 feet), Mt. Seattle (10,000 feet) and a multitude of other great peaks, many of them unnamed, still nearer at hand, while in the foreground are innumerable lesser peaks.

Causes of the Existing Glaciation. The region between Cross Sound and Cook Inlet is the seat of the most extensive present-day glaciation on the North American mainland, and includes some of the largest glaciers in the world outside of the Arctic and Antarctic regions. Yakutat Bay lies near the center of this area of extensive glaciation, because it is about in the center of the lofty St. Elias Range up whose slopes the winds, which blow prevailingly from the ocean, bear their burden of vapor day after day. The unbroken mountain chain, which causes the air to rise, interposing a lofty barrier of cold, insures a heavy precipitation of snow at all seasons. At Sitka, to the southeast, the precipitation ranges from 59 to 140 inches a year; and at Nuchek, to the west, an annual precipitation of 190 inches has been recorded. The heaviest precipitation is in autumn and winter. Both of these stations are at sea level and we have no records nearer Yakutat Bay in the mountains of the St. Elias Range; we know only that the snowfall there is exceptionally heavy, under the favorable combination of an unbroken mountain barrier and prevailing onshore winds, blowing from a sea whose temperature is raised by the inflow of warm water from the south.

On all the mountains except the very lowest there are extensive snowfields, on all of them the snowfall is heavy, and above 4000 or 5000 feet the precipitation, both summer and winter, is practically all in the form of snow. From these vast snowfields, and from others farther back in the mountains, ice tongues radiate, flooding all the valleys and filling the larger ones with vast glaciers, some of which end in the sea, others on the land at the mountain base.

General Characteristics of the Glaciers. As yet there has been no detailed study of the glaciers among the mountains back from the sea, but glimpses of portions of the region have been obtained from elevated points by various observers. The authors have looked into this snow and ice-covered mountain region from several of the low mountains near the coast, and they have had descriptions of portions of it from prospectors. The topographers of the Boundary Commission, notably A. J. Brabazon, in 1895, and the Boundary Survey parties in 1906 and 1907 in charge of Fremont Morse of the United States Coast and Geodetic Survey have seen a much wider area and have portrayed it on the maps, while some of their photographs show the condition very clearly in the snow and ice-flooded region along this part of the Alaska-Canada boundary. From a far better vantage point than either of these, upon the higher slopes of Mt. St. Elias, both Prof. I. C. Russell and the Duke of Abruzzi have looked down upon this vast sea of snow and ice. No better word picture of the supply region for the glaciers of Yakutat Bay and vicinity has been given than that of Professor Russell, which is so vivid that it deserves quotation in all descriptions of this land of glaciers. Standing upon Russell Col, on a northern shoulder near the summit of Mt. St. Elias in 1891, and looking northward, Professor Russell saw a view which he describes in the following words:

"What met my astonished gaze was a vast snow-covered region, limitless in expanse, through which hundreds and perhaps thousands of barren, angular mountain peaks projected. There was not a stream, not a lake, and not a vestige of vegetation of any kind in sight. A more desolate or more utterly lifeless land one never beheld. Vast, smooth snow surfaces without crevasses stretched away to limitless distances, broken only by jagged and angular mountain peaks. . . . The view to the north called to mind the pictures given by Arctic explorers of the borders of the great Greenland ice sheet, where

rocky islands, known as 'nunataks,' alone break the monotony of the boundless sea of ice.[1]"

Filippo de Filippi describes the same view, as seen by the Duke of Abruzzi's party from Russell Col, as follows.[2] Looking to the northwest they beheld "an interminable stretch of snow and ice, an infinite series of low mountain chains, bristling with numberless, jagged, sharp-pointed, and precipitous peaks, where rocks and ice fields were closely intermingled." From the summit of Mt. St. Elias the panorama revealed "an unexplored waste of glaciers and mountains, a vast zone bristling with sharp peaks and crags, rugged and precipitous to the south, snow-covered to the north, and surrounded by vast snowfields free from crevasses, and connected with each other by the snowy cols of the mountain chains. The medium altitude of the snow fields is about 7000 feet, that of the mountains from 9000 to 10,000 feet. No words can express the utter desolation of this immeasurable waste of ice, which Russell has compared with the ice sheet that covers Greenland. No smallest trace of vegetation can be discerned on it, no running water, no lake. It might be a tract of primitive chaos untouched by the harmonizing forces of nature. Surveying this strange scene we realized for the first time that we were close to the limits of the mysterious Polar world."

This is the condition not only north of Mt. St. Elias, but east and southeast of it (Pl. II) past the glaciers that enter Yakutat Bay, past those that enter the Alsek River, and down to the supply ground of the Muir Glacier and the glaciers of the Fairweather Range.

Where not too steep, the mountain slopes and tops are buried beneath fields of snow, and with every snowstorm more is added to be avalanched into the valleys. So much snow is sliding from the mountain slopes, or falling in the valleys, and there being transformed into glacier ice, that the valleys are drowned in a flood of snow and ice which rises high up on the valley sides. The cirques are deeply filled, the divides are buried to a great depth, and lower peaks are partly or completely submerged beneath the vast accumulation of snow and ice. The result is that there are extensive areas of snow-covered ice between the higher peaks, and these areas, being often in divide regions, are the supply ground of glaciers radiating in several directions. It is, therefore, possible to travel from the terminus of one of the glaciers up to the ice-submerged divide area, and thence proceed down to the terminus of any one of the several glaciers fed from the same broad reservoir; but it would be difficult to tell where the one glacier ends and the other begins. To such a glacier condition the name *through glacier* has been applied (Pls. III, VII, LVIII, LX, A, and Fig. 12). Where well developed, as it is east of Russell Fiord, the through glacier condition forms an intricate maze of broad, rather flat-topped glaciers of moderate slope, which so submerge the mountains as to give them the appearance of a drowned mountainous land, like the island-skirted coast of southeastern Alaska.[3]

The snowfall in these broad reservoir areas is so great that the ice surface is leveled by it, and usually the moderate slope and the deep snow-cover combine to cause a smooth

[1] Russell, I. C., Second Expedition to Mt. St. Elias: Thirteenth Ann. Rept., U. S. Geol. Survey, pt. 2, 1892, p. 47.

[2] The Ascent of Mt. St. Elias, Alaska, by H. R. H. Luigi Amedeo di Savoia, Duke of the Abruzzi, Narrated by Filippo de Filippi, illustrated by Vittorio Sella and translated by Signora Linda Villari with the author's supervision, London, 1900, pp. 148–149, 159.

[3] Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, p. 36.

surface, with few crevasses. Therefore, once the crevassed ice that is often present below the snow line is passed, these through glaciers form excellent highways. Those to the east of Yakutat Bay were followed by many prospectors, in 1898–1899, as the most feasible route to the Alsek valley, and their sledges and other relics of this over-ice travel may still be seen on and near the terminus of Nunatak Glacier, Fourth Glacier, and at the head of Russell Fiord.

Feeding the large through glaciers, both in the ice-submerged divide areas and in their lower courses, are many valley glacier tributaries which descend from the mountains that rise above the ice flood. None of them are as yet named, and none of them have so far been studied, excepting only those on the slopes of Mt. St. Elias. They appear to offer no unusual features, and nothing is known that distinguishes them from ordinary Alpine glaciers above snow line. Some are small and steep, others long and with moderate slope, and many increase in steepness near their lower ends, frequently forming ice cascades as they descend from their valleys to join the main ice mass. The latter condition, characteristic of the majority of tributary glaciers in this region, is due to former glacial erosion which lowered the main valley and left the tributary valleys hanging; for great though the expanse of snow and ice in this region is at present, it is far less than formerly existed, even in the supply areas. Indeed, in some cases, it is evident that present-day through glaciers, now flowing in opposite directions, were formerly continuous glaciers, flowing in one direction across the divide. It is, in part at least, as a result of this former condition that the divide areas have been so lowered by glacial erosion that they are now flat enough to be drowned by ice and snow and serve as the distributing area of a system of radiating glaciers. At present there is a network of glaciers flowing in various directions in the depressions between the higher peaks; formerly this network was even more extensive but the direction of flow was often changed and even reversed.

The ice supply of these through glaciers comes from three principal sources, (1) directly from snowfall, since the snow line lies between 2000 and 3000 feet in the area of the through glaciers, (2) from avalanches, (3) from tributary glaciers. From the last two sources are supplied rock fragments as well as snow, and these are incorporated throughout the ice, though naturally in greatest abundance near the glacier margins and along bands medially placed by the junction of glaciers. A third source of rock fragments is through erosion by the ice itself, and this débris is mainly concentrated in the bottom layers. From these several sources the Alaskan glaciers receive a large supply of débris; but this is not apparent above the snow line where the white, snow-covered surface (Pl. IV) shows the presence of rock fragments only occasionally where a freshly fallen avalanche has not yet been buried beneath snow. Below the snow line, however, the presence of abundant rock fragments becomes apparent, where, by ablation, they become concentrated on the ice surface, often completely covering the glacier in its lower portion.

The abundant supply of rock fragments from avalanches is dependent upon at least five favorable conditions, as follows: (1) weathering, which is very active in these high mountains, whose steeper slopes are exposed to sharp changes in temperature; (2) the abundant snowfall which, forming deep accumulations on the mountain slopes, and attaining a condition of instability, is all the time sliding down the slopes and by its erosion tearing off portions of the mountain itself; (3) the condition of the rock, which in

many places is friable schist or slate, or sheeted and jointed gneiss and granite that is readily broken off by weathering or by avalanche erosion; (4) the oversteeping of the valley slopes by long continued glacial erosion, causing cliffs and precipices down which rock fragments are all the time avalanching to the glacier surface; (5) earthquake shattering and shaking, forming rifts in the rock and thus aiding in weathering and in the production of rock avalanches, and earthquake shaking acting directly as a cause for abundant avalanches of snow and rock.

Since the St. Elias Range rises abruptly from near sea level, presenting a steep face toward the sea, the larger glaciers to which the heavy snowfall gives rise easily find their way down to or near sea level, for the snow line even on the seaward face of the mountains is not more than 3000 or 3500 feet, while in sections less exposed to the warm air and rains of the ocean it is no higher than 2000 or 2500 feet. For great glaciers to push their termini down 3000 or 3500 feet below snow line is not difficult. Shorter glaciers fail to do so, and in the Yakutat Bay region there are scores of such glaciers, some ending almost at snow line, others descending part way to sea level. Such glaciers differ but slightly, if at all, from the type of ordinary Alpine glaciers, and in a region of abundant great glaciers attract so little attention, that they have so far received little study, and are for the most part still unnamed.

Of the large glaciers which reach nearly or quite down to sea level there are four distinct types. The first of these, illustrated by Hidden, Fourth, and Black Glaciers, is not greatly different from ordinary valley glaciers, though in the case of the first two, forming a part of a through glacier system. The glaciers of this type are enclosed and terminate in mountain valleys, ending near sea level. The second type of glacier is the tidal glacier, illustrated here by Nunatak (Pl. V), Turner and Hubbard Glaciers. These may all be parts of a through glacier system, as Nunatak and Hubbard Glaciers certainly are; but below the supply area they have the characteristics of valley glaciers of large size, moving with great rapidity. Instead of ending on the land, however, they terminate in ice cliffs rising out of the sea, and discharging icebergs into it. The third type of glacier, which may or may not be part of a through glacier system, has characteristics of an Alpine valley glacier in its mountain valley, but, being forced beyond the mountain front, terminates in an expanded foot, or bulb, at the mountain base, where no longer confined between mountain walls. Such a piedmont ice bulb may lie entirely outside the mountains, as on the coastal plain which skirts the St. Elias Range, illustrated by Galiano Glacier and other large glaciers to the southeast of Yakutat Bay (Pl. I, B), or it may lie in a broader valley within the mountains, as in the case of Variegated Glacier in Yakutat Bay; or better still in the Allen and Miles Glaciers of the Copper River valley. The fourth type is the piedmont ice plateau, of which the Malaspina Glacier is the typical example, made by the union of two or more piedmont ice bulbs. Malaspina and Bering Glaciers, the largest known examples of this class, are each made by the union of many glaciers; but there are other cases of the same type where only two or three glacier bulbs are united,—for example, the piedmont bulb made by the union of Atrevida and Lucia Glaciers, just west of Yakutat Bay.

In addition to the differences between these four types of glaciers just mentioned there are other notable peculiarities. The tidal glacier maintains activity to its very front, and in all cases observed in this region is crevassed down to the ice cliff; but the other types gradually diminish in activity toward their ends and finally assume almost com-

plete stagnation, and in the case of portions of the piedmont bulbs actual stagnation. In some cases the outer portions of the bulbs become completely motionless. Moreover, the tidal glaciers terminate farther back in the mountains than the other types, for the sea eats into them more rapidly than the rains and warm air can melt away those glaciers which end on the land.

As a result of these peculiarities there are among these glaciers distinct differences in the phenomena attending the destruction of their outer portions. From the steep faces of the crevassed tidal glaciers icebergs are constantly thundering as they fall into the sea or rise from the submerged base of the ice cliff. Both ice and débris are borne away from the glacier front, the ice to melt and the débris to be deposited at a distance. Being back among the mountains, in a cooler climate, there is less melting of the surface of the tidal glaciers than of those that project as piedmont bulbs beyond the mountain face. There is consequently less water emerging from these tidal glaciers, and that which does issue comes out mainly from ice tunnels beneath the sea. Because of the rapidity of ice motion, the relative slowness of melting, and the active discharge of icebergs, the tidal glaciers have small amounts of débris upon their surfaces. Only along the lateral margins, where the motion diminishes, where iceberg discharge is either checked or entirely stopped, and where the débris supply is great, are there extensive fields of rock fragments accumulated on the surface as a result of ablation.

Those glaciers which terminate in their mountain valleys are sometimes covered by fields of débris, but in other cases have little. The Black Glacier, for example, is completely buried beneath morainic débris in the lower third of its valley portion; Fourth Glacier has some débris, but not much, while Hidden Glacier is one of the freest from débris cover of all the glaciers of the region; but this is in part, if not entirely, due to the fact that its real terminus was buried (up to 1906) beneath an outwash gravel plain. The piedmont bulbs and piedmont glaciers, on the other hand, are always either partly or completely buried beneath broad fields of morainic waste. This is due to the fact that these are large glaciers, with much incorporated rock débris, which spread out beyond the mountain front, assume a condition of stagnation or semi-stagnation, and then slowly waste away in the cool temperature, rainy climate of the Alaskan coast. Ablation then concentrates rock fragments on the ice surface, giving rise in some cases to areas of many square miles of barren rocky débris, which has been called an *ablation moraine.*[1] It consists mainly of débris avalanched down upon the glacier within the mountains, incorporated in the ice mass, and near the glacier terminus concentrated at the surface by the wasting, or ablation, of the ice. In the more stagnant portions of the glacier the ablation moraine becomes so thick that it serves to so protect the glacier that further melting proceeds very slowly. Then the soil assumes such stability that plants may find a foothold. On the inner margins of this portion of an ablation moraine willows and alder bushes are found as scattered individuals or in clusters; but on the outer margin the ablation moraine may become so stable that forests of cottonwood and spruce grow, with trees a half century or more old.[2] The Malaspina, Lucia, and Atrevida Glaciers each furnish illustrations of this condition.

[1] Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, p. 37; Tarr, R. S., Some Phenomena of the Glacier Margins in the Yakutat Bay Region, Alaska, Zeitschrift für Gletscherkunde, Band III, 1908, pp. 85–88.

[2] Russell, I. C., 13th Ann. Rept., U. S. Geol. Survey, 1894, Pls. XIII, XIV.

It is partly the presence of the ablation moraine on these piedmont glacier bulbs that permits them to remain in a temperate climate so far from the mountain front and essentially at sea level. Without the protection from the morainic cover these bulbs would of necessity waste away; with it they linger for scores of years even when all further ice supply ceases. How slowly the ice wastes is attested by the fact that dense forests thrive in the morainic soil that covers the glacier.

When the ice melts entirely away the ablation moraine forms a thin, irregular veneer on the surface where the glacier bulb formerly stood, and the sites of lateral moraines are marked by benches and terraces of morainic drift; but terminal moraines are not extensively developed at the front of the piedmont ice bulbs, though they are developed at the fronts of glaciers which terminate within their mountain valleys, and probably also in front of those which end as tidal glaciers. The failure to develop terminal moraines around the front of the piedmont bulbs is easily understood when one considers that the ice in them is either stagnant or so slowly moving that very little débris is brought to the glacier end, and that, therefore, there is no opportunity for notable concentration of deposit.

Instead of moraines around the periphery of these piedmont bulbs, the frontal accumulations that are most noteworthy are those brought and deposited by streams. There is much melting over the entire bulb area and even in the mountain valley above it, and this abundant water is supplied with great quantities of sediment. Streams issue from various portions of the ice front, some of them very large, and all heavily charged with sediment, varying in texture from fine clay to large pebbles and small bowlders. Much of this sediment must be quickly deposited on the plains that intervene between the ice front and the sea and that are themselves the product of this same stream deposit. If the ice of the piedmont glaciers of the Yakutat Bay region should melt away, without further change, the position of their fronts would be traced by a series of alluvial fans sloping away from the ice, while on the site of the ice itself there would be a thin, hummocky deposit of bowlder strewn moraine derived from the ablation moraine of the glaciers. Stream deposition is exceedingly active in this region of abundant water supply and sediment load; but moraine deposit is slow and quite ineffective.

The Glaciers of Yakutat Bay. As has been stated, there are scores of glaciers in and near Yakutat Bay, but the great majority have as yet received no detailed study, because they are small individuals in a region of much larger glaciers, and because they present no features of exceptional interest, in a region where a number of the glaciers have features of decided interest. Some of these smaller glaciers were studied and photographed during the expeditions of 1905 and 1906 and are briefly described in the reports upon the work of those years, but in the National Geographic Society's expeditions of 1909–10 attention was paid to only the larger glaciers, and those smaller ones that were of special interest. In later chapters the results of these studies, in the light of our previous knowledge of them, are stated in detail; here it is intended merely to briefly characterize and locate these more important glaciers.

Beginning with the western portion of the area, there is the Malaspina Glacier, the largest of all, a vast piedmont ice plateau made by the union of the piedmont bulbs of several large glaciers and many smaller ones. Most of its periphery is covered by ablation moraine, and in places this moraine supports a forest of alder, cottonwood, and spruce. The easternmost tributary to the Malaspina is Hayden Glacier, which, however, contri-

butes little ice to the plateau; but just to the west of the Hayden the great Marvine Glacier descends from the mountains and supplies the ice which forms the easternmost of the four lobes of the Malaspina Glacier. The low ice cliff of this lobe of the glacier lies just back of the west coast of Yakutat Bay from near Point Manby to the Kwik River, being separated from the sea by a fringe of alluvial fans across which many large, swift, glacial streams flow. The Marvine lobe of the Malaspina Glacier is of distinct present interest because of the rapid change from stagnation to activity which was observed in 1906.

East of the Malaspina Glacier, and between it and Yakutat Bay, are three glaciers which extend beyond their mountain valleys and spread out in piedmont bulbs. The largest of these, the Lucia, is the westernmost and it is now separated from Malaspina Glacier, to which it was undoubtedly formerly a tributary, only by the gravels of the Kwik River alluvial fan. Immediately east of the Lucia, and coalescing with it, is the piedmont bulb of Atrevida Glacier. Both of these bulb glaciers are partly covered with ablation moraine and on their outer stagnant termini support forests of alder, cottonwood and spruce. Atrevida Glacier changed from a stagnant to an active condition between September, 1905, and June, 1906, and Lucia Glacier was changing in 1909. Galiano Glacier, the smallest of these three glaciers, changed from stagnation to activity sometime between 1890 and 1905, probably after 1899. Its piedmont bulb extends practically to the shores of Yakutat Bay from which it is separated only by a narrow gravel beach. Two or three miles to the east of Galiano Glacier is the still smaller Black Glacier, which has no piedmont bulb, and is interesting especially because, though so near the Galiano, it gives no evidence of having undergone notable change in condition.

On the west side of Disenchantment Bay is the large Turner Glacier, a tidal glacier with an ice cliff 2½ miles in length, which, though changed slightly each time it has been observed, shows no such pronounced variation in condition as those just mentioned. Just north of it, however, is a smaller glacier, called Haenke Glacier, which, like the Atrevida, was absolutely transformed between 1905 and 1906, having become broken and having advanced nearly a mile and assumed tidal condition in an interval of ten months. Just north of this is the small Miller Glacier, which had a similar transformation about 1901.

Next to these is the Hubbard Glacier, the largest tidal glacier in the region, fed by two large tributaries from some unknown source far back among the mountains and having a tidal front 5½ to 6 miles in length. It presents many interesting features, and in 1909 had a very slight advance. Variegated Glacier, whose piedmont ice bulb coalesces with the southeastern side of the Hubbard, presents the interesting condition of a piedmont bulb in a valley, instead of at the base of the mountain front. It resembles Atrevida and Lucia Glaciers in its ablation moraine, though it lacks forest growth. It rivals Atrevida Glacier in the extent of its transformation between 1905 and 1906. Almost coalescing with the Variegated is the Orange Glacier, entirely confined to its mountain valley, unchanged since first observed by us in 1905, and forming the western end of a through glacier, whose other end is just back from the shores of Nunatak Fiord. Near the southeastern end of this through glacier, a hitherto unnamed glacier, to which we have given the name Butler Glacier,[1] descends from the mountains, and

[1] After Mr. B. S. Butler of the U. S. Geol. Survey, whose efficient aid contributed greatly to the success of the 1905 and 1906 expeditions to Yakutat Bay.

emerges from its broad mountain valley, then spreads out in a piedmont bulb reaching almost to the shores of Nunatak Fiord.

Just east of this piedmont bulb is the ice cliff of the tidal Nunatak Glacier whose history since first seen in 1891 up to 1910 was one of continued recession; but between August, 1909, and June, 1910, it advanced about 1000 feet. It has also a wasting land tongue, or distributary, and above its end hangs the Cascading Glacier, the type of a series of similar cascading glaciers in this region and elsewhere in Alaska. Opposite, on the north side of the fiord is Hanging Glacier which no longer cascades over its hanging valley lip. Hidden Glacier, to the southwest of Nunatak Glacier, was of peculiar interest in 1905 because the outwash gravel plain which separated its stagnant terminus from the sea rested for a distance on the glacier ice, which, by melting, gave rise to a pitted plain. All this is now destroyed, for when seen in 1909 Hidden Glacier was utterly transformed, having also undergone a spasmodic advance since last visited in 1906. The last glacier of the series studied in 1909–10 is Fourth Glacier, which terminates within its mountain valley, just back of the mountain front, being a typical valley glacier tongue of a through glacier system. It shows no sign of other recent change than that of recession.

Other names have been proposed for this glacier, notably Beasley Glacier, but we prefer the name Fourth Glacier, by which it is known locally, and because there seems no good reason for attempting to substitute another name. Besides being sanctioned by local usage, and by the United States Geographic Board, the term Fourth Glacier helps to record one of the bits of history of this region, as Malaspina, Haenke, Atrevida and other names record other historic events. In 1898–99 Yakutat Bay was visited by large numbers of prospectors who undertook to reach the reputed gold fields of the Alsek valley by an over-ice route, or else attempted to reach the Klondike district by an easier route than that over White Pass. Going up Yakutat Bay in boats they named the large glaciers in their order, First (Turner), Second (Hubbard), Third (Nunatak), and Fourth Glacier, passing by Hidden Glacier, which perhaps they did not see, without naming it. Parties crossed by both the Third and Fourth Glacier routes. It is an interesting episode in the history of this region, and in the mining history of Alaska, and it is our belief that it is best to retain this one of the names that these adventurous prospectors applied to the region.

Former Expansion of the Yakutat Bay Glaciers. Throughout the entire Yakutat Bay region the evidence is complete that all the glaciers have at a former period been far more extended than at present.[1] The period of greatest extension of the glaciers was recent in a geological sense, but was at least several centuries ago, for a mature forest grows on the deposits laid down by these expanded glaciers.

There are several lines of evidence upon which the conclusion is based that these glaciers were formerly far greater than now. In the first place, throughout the region the valleys show clear signs of pronounced glacial erosion. The valley walls are scored, grooved, polished and smoothed to elevations far above sea level, and, in those valleys in which glaciers still linger, to elevations far above the surfaces of the present glaciers. Tributary valleys hang above the level of Yakutat Bay, Disenchantment Bay, Russell

[1] Tarr, R. S. and Martin, Lawrence, Glaciers and Glaciation of Yakutat Bay, Alaska, Bull. Am. Geog. Soc., Vol. XXXVIII, 1906, pp. 155–164.

Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 96–137.

Fiord, and Nunatak Fiord; secondary tributaries to these lateral valleys hang above them; and hanging valleys, many with cascading glaciers, lie above the level of the surfaces of all the larger existing glaciers. Many of these glaciers head in cirques, except in the case of the through glaciers. A second evidence of former expansion is the presence of outwash gravels along the shores of the inlet in places where glaciers are no longer present or depositing gravels, and even as far out as the mouth of Yakutat Bay. The third proof is the distribution of transported rock fragments and the development of morainic terraces at elevations high above the level of the inlet, and, where glaciers are present, high above their present surfaces. Such deposits occur all along the shores of the inlet, to the west of Yakutat Bay above the eastern margin of Malaspina Glacier, and in the valleys of the larger glaciers which come down to Yakutat Bay. The moraine terraces descend in the direction of the ocean, and are evidently to be correlated with the deposits which constitute the fourth evidence of former glacier expansion,—the hummocky moraine which forms the southeastern margin of Yakutat Bay, as far out as Yakutat, the similar moraine about the head of Russell Fiord, and the crescentic deposit which extends as a submarine ridge across the outer part of Yakutat Bay.

From these four lines of evidence it has been concluded that, at the period of greatest expansion, all the glaciers were much larger than now,—Malaspina Glacier then rose much higher on the slopes of the mountains west of Yakutat Bay, its tributaries were greater, it received tributaries, notably Lucia and Atrevida Glaciers, that are now disconnected, and it coalesced with a great glacier that filled Disenchantment Bay and Yakutat Bay out as far as Yakutat and the submerged moraine that stretches in crescentic form from Ocean Cape westward to Point Manby. To this expanded glacier that filled Yakutat Bay, the name Yakutat Bay Glacier has been given; and the similar expanded glacier in Russell Fiord has been called the Russell Fiord Glacier. The latter glacier completely filled Russell Fiord and terminated in a piedmont bulb on the inner edge of the foreland, where it has left a crescentic moraine from which outwash gravels slope seaward.

Since the period of maximum glacier expansion, and far more recently than it, there has been a minor advance, by which the united Hubbard and Turner Glaciers, joined by others, pushed southward into Disenchantment Bay and southeastward into Russell Fiord, while Nunatak Glacier, coalesced with Hidden Glacier and others, pushed northwestward into the northwest arm of Russell Fiord, and southward into the south arm about two thirds of the way to the head of the bay. During this advance a lake was formed in the upper end of Russell Fiord whose shoreline can be easily seen and traced. This advance of the glaciers was of such brief duration, and of such moderate intensity, that the ice erosion did not succeed in removing the gravels previously deposited. It, therefore, contrasts strikingly with the earlier, prolonged advance by which the bed rock was scoured out to a depth of many hundred feet by the powerful erosive action of the expanded glaciers. Between these two ice advances there was a long interval during which the glaciers receded even farther than at present, and forest growth extended throughout the fiord, and even up the valleys now occupied by the glaciers. The termination of the last advance was but a short time ago, at the most but a century or two, and the recession from this stage of advance was apparently still in progress as late as 1905. The proof of the recency of the last advance, and of the ice recession from that stand, is the condition of the vegetation growing in the area occupied by the ice and the

lake water. In the outer portion of the area covered by the expanded glacier, a dense growth of mature alder and some cottonwood covers the overridden gravels, but the growth rapidly decreases in amount and density toward the present glacier fronts. In Seal Bay and Nunatak Fiord there are only scattered individual plants, and the density of alder growth gradually increases thence toward the portions of the inlet where the expanded glaciers ended. In other words, this period of ice advance was so recent that only a part of the area is as yet occupied by vegetation (Pl. VI), and the outer portion is occupied only by the advance growth of alder and, in the extreme south, of cottonwood. The spruce forest of the Alaskan coast has not yet had time to advance upon the region from which the glaciers have so recently receded.

Recent Recession of the Glaciers. In Professor Russell's visits to the Yakutat Bay region in 1890 and 1891 he found the glaciers to be in general in a state of recession. Dr. Gilbert's observations in 1899 led him to the same conclusion, and our observations in 1905 showed that the glaciers were still wasting away. The evidence of this condition of recession is partly inferred from the characteristics of the glaciers and the conditions at their borders, and partly observed by direct comparison at the later dates with observations made during the earlier studies. Russell, Gilbert, and the authors of this book, have all noted the fact that many of the glaciers are covered by ablation moraines at their lower ends, and that in some of the more stagnant portions these ablation moraines bear forests. From this condition the inference is perfectly warranted that the glaciers in these regions are wasting. More specific, however, is the evidence around the glacier borders, both at the sides, and, in those which end on the land, at their fronts. While forest, or at least alder (Pl. VI, B), extends nearly up to the front of many of the glaciers, and also grows on the valley sides above them, there is, near many of the glaciers, a zone in front of and just above the glaciers, which is barren of vegetation (Pl. VI, A). From such a condition one infers with certainty that the ice has withdrawn from such areas so recently that vegetation has not yet had time to encroach upon it. The extent of shrinkage indicated by this class of evidence varies with different glaciers, but it was present to some degree in the neighborhood of almost all the glaciers studied, and in some it indicates a great and long-continued shrinkage. This is particularly true in Nunatak and Russell Fiords, as already stated in the preceding section. Here it is certain that in the last century the recession has amounted to many miles.

By the photographs taken by Professor Russell, though they were taken merely as incidents in a work of different object, a basis was established for future record of the changes in the position of the fronts of some of the glaciers; and still further basis for such comparative work was established by the photographs of the Canadian Boundary Commission in 1895. Naturally, therefore, Dr. Gilbert instituted such comparisons when, in 1899, he visited the region as a member of the Harriman Expedition; and he added a new basis for future comparison by taking photographs of still other glaciers. We, in 1905, therefore, had a body of photographic evidence upon which to study the changes of positions of the glacier fronts during recent years. This study of photographic records shows that some of the glaciers, notably Hubbard and Turner, changed but little between 1891 and 1905, but that Nunatak and Hidden Glaciers receded greatly between 1899 and 1905. For the smaller glaciers there is no photographic basis for a statement, though the other evidence gives conclusive proof of recession in many cases.

In 1899 the Harriman Expedition added still another basis for future studies of the

positions of the larger glaciers by the construction of plane table maps by Mr. Henry Gannett. Maps of similar character were also made by the United States Geological Survey expeditions of 1905 and 1906 by Messrs. Butler (in 1905) and Rich (in 1906). It was part of the plan of the National Geographic Society's expeditions of 1909–10 to make more detailed maps of the larger glaciers of Yakutat Bay as a basis for future studies of the changes in these glaciers, and for this purpose one of the topographers of the United States Geological Survey, Mr. W. B. Lewis, was attached to the expeditions. Assisted by E. F. Bean and F. E. Williams, he has made the contour maps which accompany this book. All future students of the glaciers of this region will find in their carefully-made maps a reliable basis for comparison of the positions of the glacier fronts as they may change from time to time.

Recent Advance of the Glaciers. Earlier in this chapter it has been intimated, and in other publications stated with full detail,[1] that some of the glaciers in the Yakutat Bay region have been recently subjected to an advance of such unusual character as to lead to the desire to study the phenomenon further. It was primarily this desire that led us to return to the Yakutat Bay region in 1909 and 1910, and the results of the work of these years that are of most value relate to the phenomena attending the advance of the glaciers. These results will be stated in detail in later chapters devoted to the discussion of individual glaciers; but in order that these details may appear in their proper setting it seems desirable to make a general statement concerning the phenomenon.

In 1905, while a general condition of recession characterized the great majority of the Yakutat Bay glaciers one, Galiano Glacier, presented convincing evidence of change to activity in the interval since it was photographed by Russell in 1890. Then it had a stagnant piedmont bulb on whose ablation moraine a forest of alder and cottonwood grew, proved both by Russell's description and by his photographs and also shown by Boundary Survey photographs in 1895. This forest growth had entirely disappeared in 1905, but the piedmont bulb was again stagnant and covered by ablation moraine, though with only young alders scattered here and there. Neighboring glaciers, for instance Atrevida to the west and Black Glacier to the east, gave no evidence of similar change, and no such evidence was found in any other glaciers. Puzzled by the phenomena, and forced to the hypothesis that it was in some way connected with the earthquakes of 1899, we made no attempt at explanation but merely described the facts. We were not at all prepared for the momentous changes which occurred in the interval between the summers of 1905 and 1906.

Returning to Yakutat Bay in 1906, the senior author found four glaciers absolutely transformed, and all the others unchanged. These glaciers that were so altered in the brief interval of nine or ten months are, named from west to east, the Marvine Glacier and the eastern lobe of the Malaspina Glacier that is fed by the Marvine, the Atrevida Glacier, Haenke Glacier, and Variegated Glacier. In the summer of 1905 one could travel over the surfaces of these glaciers at will. On two of them, Atrevida and Variegated Glaciers, we walked freely, on the former late in August, without recognizing any signs of coming change to activity. The glaciers were crevassed slightly only here and there, and outside the mountain were in a stagnant or semi-stagnant condition and covered with a waste of ablation moraine; but in June, 1906, all four glaciers were trans-

[1] See especially, Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 41–95. Here will be found references to other publications on the subject.

formed to a sea of crevasses; and not only was it impossible to travel over their surfaces, but it was not even possible to climb up on the glaciers except by the most difficult ice work. Furthermore, the glaciers were even then actively advancing, and the advance extended out even to the fully-stagnant margins, overturning forests of alder and cottonwood that were growing on the outer portions of the Malaspina and Atrevida Glaciers. Not only were the ice surfaces broken by a maze of crevasses, and the margins which had hitherto been gently-sloping, moraine-covered icebanks transformed to steep ice cliffs crowned by bristling ice pinnacles, but the margins were pushed forward, and the hithertofore stagnant piedmont bulbs were thickened. Haenke Glacier had advanced to tidal condition; the Atrevida and Malaspina Glaciers were pushing into the forest that fringed their margins; and Variegated Glacier had become notably thicker and had crowded out over a rock gorge, destroying the glacial stream that had occupied it in 1905.

It was evident from these facts that the glaciers in question had been subjected to some unusual impulse that had caused a sudden forward movement by which they had been pushed forward, thickened, and their surfaces greatly broken. In seeking for an explanation for such a phenomenon, not hitherto recorded, only one cause seemed adequate, namely, the effect of the severe earthquake shocks to which this region was subjected in 1899. The hypothesis put forward by the senior author was that the repeated violent shaking during the earthquakes that occurred between September 3 and 29, 1899, threw so much snow and ice into the reservoirs of the glaciers that a wave of motion was started which reached completely down to the terminus of Galiano Glacier before 1905, and which was passing down the four other glaciers during 1906. In testing this hypothesis with the facts available, all were found to be in harmony with it, none were discovered that were opposed to it, and no other hypothesis could be suggested which had no fatal objections to it.

While the hypothesis of earthquake cause for this advance seemed, therefore, well supported it was desired to subject it to still further test, and it was one of the main objects of the expedition of 1909 to apply these tests. There were three such tests which we had especially in mind. In the first place, if the advance were due to this cause, it should be confined to the region of violent earthquake shaking. By inquiring regarding the condition of glaciers southeast of Yakutat Bay, as in Glacier Bay, and by study of some of the glaciers of the Prince William Sound region to the northwest we were able to apply this test to some extent, but not so fully as to warrant definite statement of its adequacy in support of the hypothesis. In view of the multitude of glaciers in the region to the southeast and northwest of Yakutat Bay, the fact that among the comparatively few which we were able to study or gain specific information about, we found none that furnished evidence of spasmodic advance, is only contributory evidence. For final testimony other studies over a wider field are necessary.

The second test is that of the behavior of other glaciers in the Yakutat Bay region in the interval between 1906 and 1909. If the hypothesis proposed is correct, probably some of the smaller glaciers of Yakutat Bay had advanced before 1905, and certainly some of the other glaciers of the region ought later to show signs of the appearance of the wave of advance. This was predicted in the report on the expeditions of 1905 and 1906.[1] There is reason to believe that there was an advance of some of the smaller glaciers before 1905, though it is now difficult to obtain convincing evidence; but that the

[1] Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 93–94.

wave of advance was extended to other glaciers between 1905 and 1910 was strikingly illustrated by Hidden Glacier, which had advanced two miles and become greatly broken, by Lucia Glacier, which was actively advancing and greatly transformed by crevassing during the summer of 1909, and less strikingly by Hubbard Glacier, whose eastern margin was just beginning to advance in 1909, and by Nunatak Glacier, which had a slight advance in 1910.

The third test was that of cessation of advance in those glaciers that had moved forward in 1906. With a sudden, great addition of snow, quickly terminated, and followed by a spasmodic advance of the glaciers thus supplied, it would be expected that the wave of advance would soon die out and the condition of stagnation return. This was also predicted,[1] and our observations of 1909–10 show clearly that the prediction was correct, for all of the advancing glaciers of 1906 had returned to stagnation in 1909, and the breaking of the ice had been so healed by ablation that we were once more able to travel over the glaciers, though with far less ease than in 1905.

It is believed that the observations of 1909 add the further facts necessary to demonstrate the hypothesis put forward in 1906, and that we are now warranted in stating the explanation with confidence, as an established hypothesis,—a new cause for glacier advance. The sudden forward rush of a glacier, accompanied by pronounced thickening and extensive surface breakage, may be called a *glacier flood* and the resemblance to a river flood is noteworthy. When heavy rainfall, or unusual melting of snow occurs in the upper reaches of a river, a wave of rising, rapidly down-moving water is started which may cause a flood all along the course, and if a portion of the river is ice covered, the rigid ice crust will be shattered and heaved into a maze of broken ice blocks; but under ordinary conditions the river behaves more normally, slowly rising and falling with variation in supply. So in a glacier, under ordinary conditions, variations in supply manifest themselves in moderate advance or recession; but when a deluge of snow and ice is thrown down in its upper reaches the condition for a spectacular advance, a glacier flood, are introduced. The ice stream flows on more rapidly, its rigid upper and marginal portions are cracked and broken, its surface rises, and its front is pushed forward. There is, however, a striking difference in time occupied by the two classes of floods. A river flood passes from the source to the mouth of the river in a few hours or a few days, and its effects are soon past; but the far less mobile ice requires several years for the transmission of the glacier flood, and its duration is months long, while years are required to bring the ice surface back to its pre-flood state. There is a striking difference, also, in the mode of motion, for the mobile river water itself travels down stream to form the flood while in the glacier flood the advance and breaking of the mass is due to the passage of a wave, not to actual bodily transfer of ice from glacier reservoir to terminus.

With these facts in mind it is quite proper to predict the advance of other and longer glaciers in the Yakutat Bay region as a result of avalanching during the 1899 earthquakes. The oscillations of other Alaskan glaciers, for which there is no known climatic warrant, may be due to other earthquakes. Earthquake avalanching may even be responsible for oscillations of mountain glaciers in other regions, as for example in the Himalayas and other youthful, snow-capped mountains still frequently faulted and shaken by seismic disturbances.

[1] Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, p. 94.

With this general view of the glaciers of the Yakutat Bay region as a preliminary basis for understanding their main characteristics and their behavior since first studied by Russell in 1890 and 1891, we will proceed to the study of the individual glaciers, and then, upon this basis, to a consideration of the features of general interest and their interpretation, including their bearing upon the interpretation of the Pleistocene drift phenomena of North America and Europe. In this study free use will be made of the results of previous work, as well as of the new observations of 1909 and 1910, the purpose being to set forth the phenomena of glaciation in this region in the light of a series of investigations each of which has added something to a story of glacier condition and behavior of more than ordinary interest, and of quite unusual character.

CHAPTER III

THE MALASPINA GLACIER AND ITS TRIBUTARIES

The Malaspina Glacier

General Description. From snow-covered Mount St. Elias and Mount Logan, 18,000 and 19,500 feet high, and from the 10,000 to 16,000 foot mountains to the east and west a multitude of glaciers descend, the six largest of which named from west to east, are the Guyot, Tyndall, Agassiz, Seward, Marvine, and Hayden. These glaciers, fed by many tributaries, have their sources far back among the mountains and their upper portions are unknown.

Where they emerge from their mountain valleys they are broad ice streams, ranking among the largest of the Alaskan valley glaciers. The great volume of ice which flows out of these valleys suffices to build the broad ice plateau (Pl. VII) which skirts the mountain base from Yakutat Bay to Robinson Hills, which, according to Russell, has an area of about 1500 square miles, more than the area of the state of Rhode Island, and far the largest glacier in America, or, for that matter, in all the world, excepting only in the polar regions.

The single glaciers lose their individuality at the mountain base, where they all coalesce to form the continuous ice plateau; but they nevertheless maintain an appreciable influence on the Malaspina Glacier even to its outer margin. Because of this influence the ice plateau really consists of more or less pronounced lobes with boundaries sometimes marked by medial moraines, and with a moderately lobate front. Tyndall and Guyot Glaciers, deflected by the Chaix Hills, a south-extending spur of Mt. St. Elias, appear to unite to form a pronounced lobe on the western margin, where the ice formerly crowded out into the sea and was terminated by a magnificent cliff from which icebergs were discharged into the open Pacific. From all the photographs and descriptions which we have been able to obtain from publications of explorers and from conversation with prospectors this western, or Guyot, lobe of the Malaspina differs from the eastern portion of the glacier in maintaining a generally crevassed condition from the mountain base to the sea. It is said to present an insurmountable barrier to travel, and it is the common belief of prospectors in this region, that the Malaspina Glacier is absolutely inaccessible from the west. It is certain that no one has yet crossed it. The Tyndall Glacier was easily traversed by the Schwatka and Topham parties in 1886 and 1888 but no attempt was made to cross the Guyot.

The Agassiz lobe, next east of the Guyot, fed by Libbey and Agassiz Glaciers, is a well defined lobe, bordered by moraines on both the eastern and western margins, and faced by a broad tract of stagnant and semi-stagnant ice bearing an extensive waste of ablation moraine, and on its outer edge supporting a forest of spruce and cottonwood. Around the periphery of this lobe there is the greatest development of ablation moraine, and

apparently the broadest area of stagnation. Here too, we find the ice front farthest from the sea, and with the broadest fringe of outwash gravels; and it is from the western margin of the lobe at its junction with the Guyot lobe, that the Yahtse River emerges, one of the largest glacial streams that pour out from the Malaspina front.

Adjoining the Agassiz lobe on the east is the Seward lobe, which, so far as our present knowledge permits us to judge, is the largest of the Malaspina lobes. Certainly no other exceeds it in size excepting possibly the Guyot, about whose extent and characteristics little is known. On the western side and front it is fringed completely by a border of ablation moraine, but on the eastern side it is not definitely separated from the Marvine lobe. It is in the Seward lobe that the Malaspina Glacier extends farthest from the mountains, reaching the ocean at Sitkagi Bluffs, but not discharging icebergs. At this point there is a low, moraine-covered cliff against whose base the waves of the Pacific Ocean break (Pl. IX, A).

The easternmost lobe of the Malaspina Glacier is that dominated by the Marvine Glacier. It is far less pronounced than the others, and, as already stated, the boundary between it and the Seward lobe is not marked by any line of moraine. Since Seward Glacier is much larger than Marvine it might be inferred that the Seward lobe would crowd eastward to the shores of Yakutat Bay, and possibly it does so under normal conditions; but in 1906, when Marvine Glacier was advancing, the ice broken by the advance extended along the entire shore of Yakutat Bay almost to Point Manby. It is quite probable, however, that the western portion of the broken ice was really a part of the Seward lobe temporarily affected by the forward thrust of the Marvine ice. On its eastern side the Marvine lobe spreads out nearly to Yakutat Bay and to the Kwik River. It is enabled to spread out in that direction because of the absence of any obstacle; but toward the west it competes with the Seward lobe for an opportunity to spread out as a piedmont bulb. Really, therefore, the Seward and Marvine Glaciers combine to form one broad lobe, each glacier dominating one side of the lobe and at the contact so merging into one another that the line of division is not recognizable. The Marvine or eastern portion of this lobe is fringed by a narrower border of ablation moraine than common, and its front is separated from Yakutat Bay by only a narrow fringe of outwash gravel. The extreme eastern margin of the lobe, which forms the western border of the Kwik River valley, was, up to 1906, in such a stagnant condition that it was covered by a thick ablation moraine in which grew a spruce, cottonwood, and alder forest. This stagnant part represented that portion of the lobe that spread farthest east and was, therefore, least under the influence of the supply of ice which the Marvine Glacier contributes to make the lobe. The entire eastern margin, nearly to the mountain base, is covered with ablation moraine.

The Hayden Glacier, unlike the other tributaries, forms no great lobe, in this respect resembling many smaller glaciers that unite with the Malaspina. It joins the ice plateau, athwart the current of the Marvine, and almost at right angles to it. At Blossom Island the Hayden Glacier coalesces with the Marvine lobe of the Malaspina and prior to 1906 the drainage of the Blossom Island region escaped through a subglacial tunnel at this point; but the advance of Marvine Glacier in 1906 closed this tunnel and the drainage found a new outlet farther up the margin of the Hayden Glacier.

In 1905 it was not possible to say where the Marvine ice and Hayden Glacier joined, both being stagnant. In 1906 the contact between the Hayden and Malaspina Glacier

was clearly indicated, from Blossom Island southeastward, by the crevassed ice of the Marvine lobe, and it was evident that the Hayden Glacier contributed little to the ice supply of the Malaspina. On emerging from its mountain valley the Hayden Glacier spreads out, and in 1906 its crescentic outer boundary was shown with almost diagrammatic clearness by the change in condition between the smooth surface of the ice contributed by the Hayden and the broken, crevassed ice of the Marvine lobe. Besides pushing out into the Malaspina, the current of the Hayden ice is deflected southeastwards along the western base of the Floral Hills, giving rise to a narrow, wedge-shaped area which in 1906 was uncrevassed, in striking contrast to the ice of the Malaspina to the west and south of it. The Kwik River emerges from its subglacial tunnel at the point of contact between the Hayden wedge and the Marvine lobe.

We have no definite knowledge of the rate of motion of any part of the Malaspina Glacier but from Russell's description, and from our briefer study of it we are confident that there is great difference in rate of motion in different parts. In the crevassed Guyot lobe there must be rapid movement down to the sea; and that there is rapid motion where the great valley glaciers emerge from their mountain valleys is indicated by the crevassing there, as shown by the Abruzzi photographs of the débouchure of Seward Glacier, by the enormous ice plateau which they supply and maintain, and by areas of crevassing in the valley portions of the tributary glaciers. From these extremes there is every gradation in rate of motion to complete stagnation in interlobate areas and in those portions of lobes where the ice has spread out farthest; but throughout the greater part of the ice plateau there must be some motion in order to keep up the supply. That this motion is not rapid, however, is indicated by the general absence of extensive crevassing. In general the surface of the Malaspina Glacier during the period of observation, up to 1906, was smooth and slightly broken. There were of course some cracks, and some areas of crevassing, there were moulins, and there was roughness due to differential ablation; but that there was a condition of general smoothness, as glaciers go, is indicated by the successsful use of the Malaspina Glacier as a highway of travel and transportation of supplies, partly by sledges, by all the mountain-climbing expeditions, over several routes.

In winter the entire surface of the glacier is buried beneath a thick blanket of snow, and from much of its surface this is melted off by the first of August, but patches remain in protected places throughout the summer. The snow line on the Hayden Glacier, according to Russell, lies at an elevation of 2500 feet, which brings it well up in the mountain valley tributaries of the glacier. This agrees with our observations of 1905 when the snow line was above Floral Pass late in August, and in 1906, when we found the snow line just below Floral Pass, in July. In 1905 its elevation was above 1500 feet at least, but in 1906, owing to an accident to our barometers, we were unable to determine its elevation. While some of the higher parts of the Malaspina Glacier may rise above snow line, especially near the mountains, most of its ice surface is exposed to melting during a part of the summer. In the lower, outer portions, the summer ablation continues through several months, but on the higher portions, near the mountains, the period of ablation is very brief. Observations on Hayden Glacier between Floral Hills and Blossom Island, in late July, 1906, showed a rate of ablation of about 4 inches a day and at this point the water on the ice froze as soon as the sun sank low in the western heavens. The rate and extent of ablation on this glacier is so great that unless it were in motion throughout

most of its area it could not maintain its size and form. Back of the fringe of ablation moraine, which checks ablation, there would, of necessity, be a depressed area due to wasting away of the clear ice unless new supplies were being constantly added.

The surface of the Malaspina Glacier, is, in general, a vast expanse of clean ice (Pl. IX. B); but the monotony of the waste of ice is relieved by areas of moraine, especially prominent in two positions, as ablation moraine around the periphery (Pl. X, A) and as medial moraines between lobes. The latter areas represent lateral moraines of the glaciers forming the lobes, brought in relief by ablation. The former, consisting mainly of material incorporated in the ice in the mountain valleys and the glacier reservoirs, is concentrated at the surface in broad tracts, and in places to considerable depth, by the ablation of the ice through which the débris was scattered. Possibly some of this peripheral moraine is supplied by uprising bottom layers, but of this we have no evidence. A third class of moraine, more difficult to understand, is long, sinuous lines of scroll-like moraine, forming a black tracing in the expanse of clear white ice. Russell suggested as explanation of such moraines some unknown variations in ice currents and this is possibly a correct interpretation though it is difficult to conceive the process. Another hypothesis suggests itself to us, namely, that these sinuous moraines may, in part at least, represent avalanche falls upon the upper glaciers, spread out in lines by the currents of the ice. We have seen them only from a distance, but they are most striking phenomena, resembling the lines of foam that develop in a swirling river current.

The periphery of the Malaspina Glacier is for the most part skirted by a fringe of alluvial fan deposits, crossed by a multitude of streams of which the largest, all raging torrents, are the Kwik, Kame, Osar, Manby, Yahna, Fountain, and Yahtse (Pl. VIII). All, burdened with vast quantities of clay, sand, and gravel, are depositing rapidly in the sea and thus extending the area of the fringe, and, by deposit in their beds, being forced to divide and subdivide, and constantly shift their courses, are raising the surface of the alluvial fringe. Where the streams cross the narrow strip of outwash gravel plain the rapid deposit by the ever-shifting branches prevents the growth of vegetation and, therefore, gives rise to barren alluvial fans; but in those portions where only small streams flow there is a growth of vegetation, in places spruce and hemlock forest, elsewhere cottonwood, or only alder and willow. Because of the fact that the plain lies near sea level, while the ice immediately back of it is so covered by moraine that it has little effect on the climate, the forest here grows as luxuriantly as elsewhere on this part of the Alaskan coast wherever the absence of glacial streams gives sufficient freedom from destructive inundation. This forest grows not only in front of the glacier, but, wherever there is a sufficient degree of stagnation of the ice to permit the growth of trees without the undermining action of a slumping soil, it encroaches on the glacier itself (Pl. X, B). This is not a general condition around the whole margin of the Malaspina but is confined to two limited regions of exceptional stagnation,—along and near the Kwik River, and in the neighborhood of the Yahtse River.

Malaspina Glacier is interesting in itself as a perfect type of a kind of glacier not recognized until Russell described its characteristics. It has an even wider interest from the fact that it is an existing example of a type which in former times must have been far more extensive. During the period of former expansion of Alaskan glaciers it was duplicated in many places along the Alaskan coast. The piedmont type of glacier was repre-

sented in former times in many parts of the Rocky, Cascade and Coast Ranges of western United States and Canada when glaciers flowed down these mountain valleys to the base. The type may very well have appeared in Norway and Great Britain as the ice sheets of the Glacial Period advanced from the highlands, and again as the waning ice sheets shrank back toward the highland areas. In the Alps, too, conditions favoring the development of piedmont glaciers appeared. Indeed, wherever mountain glaciers passed beyond their mountain valleys out upon level land, expanded piedmont bulbs must have formed, as in the valleys of the Italian lakes south of the Alps; and where two or more such bulbs coalesced piedmont glaciers resulted, as upon the plateau north of the Alps. For the interpretation of the evidence of such former piedmont glaciers the study of the conditions on and around the Malaspina Glacier throw much light. The resemblance of those piedmont glaciers which have vanished to those which still exist in Alaska in the main features, both of form and deposits, must have been great, for in both cases the conditions of supply are similar, and in both cases the conditions of wastage in a cool temperate climate are alike.

Early Observations on the Malaspina. The early explorers of the eighteenth century seem to have recognized the glacial character of this great ice mass, for Vancouver, in describing Icy Bay in 1794,[1] says he saw a "high abrupt cliffy point forming the west point of a bay, bounded by a solid body of ice or frozen snow." Tebenkof's Icy Bay chart of 1848 (Fig. 1), based upon the log books of several Russian explorers sent out in 1788 and from 1793 to 1807, shows eight miles of glacier fronting on Icy Bay behind which is a surface supporting trees.[2] Belcher in 1837 [3] says that "the whole of this bay and the valley above it was now found to be composed of (apparently) snow ice about thirty feet in height at the water cliff. . . . The small bergs or reft masses of ice, forming the cliffy outlines of the bay, were veined and variegated by mud streaks like marble, and where they have been exposed to the sea were excavated into arches, etc., similar to our chalk formations. . . . In Icy Bay the apparently descending ice from the mountains to the base was in irregular broken masses." Vancouver also said, "As the vessel advanced along the coast from Point Riou the country became less woody, and beyond the low, projecting point it seemed only to produce a brownish vegetation, which further to the eastward entirely disappeared, and presented a naked, barren country, composed apparently of loose unconnected stones of different magnitudes." Tebenkof described the low country between Mt. St. Elias and the sea as a forest-covered tundra where "through cracks in the gravelly soil, ice could be seen beneath."

Possible Great Advance of the Malaspina. These statements, some of which have been quoted previously by Russell [4] and by Davidson,[5] assign to the Malaspina Glacier such different conditions from those observed since 1890 that one is led to question whether the difference can be entirely due to failure of observation. A possible explanation of the difference may be that before 1794 Malaspina Glacier had undergone a long period of stagnation and recession resulting in the ablation-moraine surface of "naked barren

[1] Vancouver, Capt. George, Voyage of Discovery, Vol. V, 1801, pp. 349, 351.

[2] Tebenkof, Capt. Michael, Hydrographic Atlas and Observations, with 48 charts. St. Petersburg, 1848 and 1852.

[3] Belcher, Capt. Sir Edward, Narrative of a Voyage Round the World, London, 1843, Vol. I, pp. 79–80.

[4] Russell, I. C., Nat. Geog. Mag., Vol. III, 1891, pp. 66–70.

[5] Davidson, George, The Glaciers of Alaska that are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc. Pacific, Vol. III, 1904, pp. 43–47, and Map VI.

country" with "loose unconnected stones" and ice beneath cracks in the gravelly soil observed by Vancouver and the Russians, with trees growing in it where the crevassed Guyot Glacier is now unencumbered, with the shore-line east of the Yahtse River shown on the 1788–1807 chart markedly different from that of Russell in 1891, and with the deep re-entrant of the Icy Bay of Vancouver, Belcher, and Tebenkof. Subsequently there may have been an advance of Malaspina Glacier front, notably altering the conditions. This advance would probably be shortly after 1837, as the surface was not apparently much crevassed when seen by Dall and Baker in 1880, by Schwatka in 1886, and Topham in 1888, and trees had grown to good size before Russell's visit in 1891. Russell neither mentions nor shows photographs of any trees on the west or central part

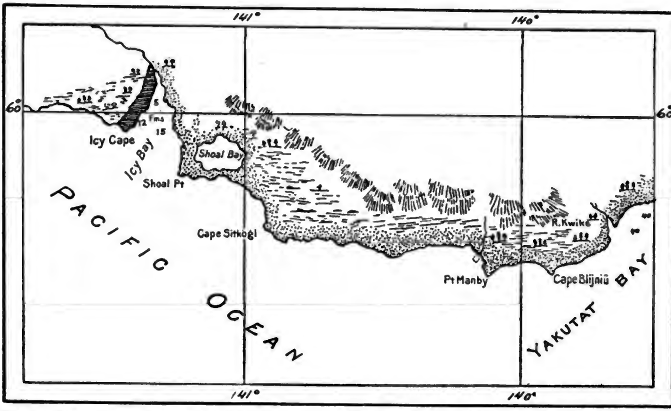

FIG. 1. TEBENKOF'S CHART SHOWING ICY BAY, ICY CAPE AND THE SITE OF THE PRESENT MALASPINA GLACIER.

of Malaspina Glacier old enough to disprove this interpretation. Davidson suggests that Malaspina Glacier has crowded its way seaward obliterating the old Icy Bay.[1]

The most specific evidence in connection with this interpretation, with which we agree, and not discussed by Davidson, is found in a map, and a native legend. Among the maps showing Icy Bay, we refer to that from Tebenkof's Atlas (reproduced here as Fig. 1). The native legend is one quoted in Topham.

The map shows a very definite Icy Bay five miles wide at the mouth, extending northward six to eight miles and apparently about on the site of the present Yahtse River. Within it are soundings of 5, 12 and 15 fathoms where only a delta existed at the time of Schwatka's, Topham's, and Russell's visits in 1886, 1888, and 1891, as their maps show. Tebenkof's chart also shows a pronounced cape just east of Icy Bay, Shoal Point (the Pt. Riou of Vancouver), behind which is a good sized lake or lagoon, Shoal Bay,

[1] Op. cit., p. 45.

which no longer exists. The shore-line eastward to Yakutat Bay is also very different from the present coast.

Russell explained the absence of Icy Bay by supposing that it had been filled by the glacial streams and that its site is the present delta of the Yahtse River.[1] That this interpretation is erroneous, and that Icy Bay was destroyed by glacial advance is indicated by the testimony of Yakutat natives. H. W. Topham was told in 1888 by George, second chief of the Yakutats, of this advance. He says:[2] "There is a tradition amongst his people, that formerly there was a large bay running up from the sea to the very foot of St. Elias; that there was a village at the head of that bay; that all around the village was swampy or muddy (Yahtsé) ground; that the mountain was, therefore, called Yahtsé-tah-shah, *tah* meaning harbor, and *shah* meaning peak; that a river flowed into the bay from the northwest, where were large glaciers; that the east of the bay was all ice, but the west, sand and trees; that at the mouth of the bay dwelt some Indians, and that one day an Indian came rushing home crying 'Quick, quick, the ice is coming,' pointing to the river down which the ice was seen to be rapidly advancing. The Indians escaped along the shore. The ice came on right across the bay, till it struck the opposite shore, when it turned and continued down the bay to the sea, swallowing the village in its course."

The description of the bay at the very foot of Mt. St. Elias, with the river coming in on the northwest, agrees with the Tebenkof map, except that the west side had the glacier and the east side sand, swamp, and trees according to the map, but Mr. Topham may very well have misunderstood the natives about this, or the native story may have been in error. Native testimony and native legends are not very trustworthy bases for scientific conclusions; but in this case there is some reason for believing in the general accuracy of the story, for it is in harmony with the map which we have reproduced. While, therefore, we cannot consider it demonstrated, we believe it possible and even probable that early in the nineteenth century the front of Malaspina Glacier has been very decidedly modified by pronounced advance. There is nothing in the present condition of vegetation in the region that disproves such an advance within the period between 1837 when Belcher "reached into Icy Bay" and "tacked in ten fathoms, mud,"[3] and 1886 when Schwatka, Libbey, and Seton Karr found no Icy Bay, but a Malaspina Glacier 25 miles wide between the south end of Chaix Hills and the ocean.[4] Indeed, Seward lobe may have possibly felt the impulse of advance before the Agassiz and Guyot lobes and have advanced before Belcher's visit, for he mentions that the small islet under Pt. Riou shown by Vancouver was no longer there.[5] The diameter of the trees on the ice near Pt. Manby, as described by Russell, suggest that the advance on this side was perhaps before 1837. The advance was pretty surely before 1874, when Dall made his first visit to Malaspina Glacier alluded to below, but as he did not visit the part of the ice front near Icy Bay this cannot be settled positively.

The establishment of a modern advance of such magnitude just west of Yakutat Bay

[1] 13th Annual Report, U. S. Geol. Survey, 1894, p. 13.

[2] Topham, H. W., A Visit to the Glaciers of Alaska and Mt. St. Elias, Proc. Roy. Geog. Soc., Vol. XI, 1889, pp. 432–433.

[3] Op. cit., Vol. I, 1843, p. 79.

[4] Seton Karr, H. W., Shores and Alps of Alaska, London, 1887, map on p. 87.

[5] Op. cit., Vol. I, 1843, p. 79.

would be very important, for it matches the latest great advance in Yakutat Bay and Russell Fiord,[1] which we have tentatively dated as not much over a century ago, but from which the ice has now retreated 15 to 25 miles, as will be described in a later chapter. That Malaspina Glacier has had a recent great advance, though of a date which cannot be stated exactly, is proved by the forest trees incorporated in the gravels at Blossom Island, although trees are growing on its summit. This advance may have occurred within a century, and it may be indicated by the evidence above stated; but we cannot correlate it with certainty.

Later Observations on the Malaspina. Between May 21 and 26, 1874, a Coast Survey party visited Yakutat Bay and made a determination of the elevation of Mt. St. Elias.[2] At this season the entire surface of the Malaspina Glacier was so covered with snow that its true character was not discovered; but in June, 1880, when the snow had disappeared from the lower portions of the glacier, the fact that it was a low-lying ice plateau at the mountain base was discovered by Dall and Baker and the name Malaspina was given to it.[3] It was then stagnant and partly covered with moraines with some vegetation on the ice near Pt. Manby. In the next decade several expeditions crossed a part of the glacier and made observations upon it, though the main object of the expeditions was the ascent of Mt. St. Elias rather than glacier study. These were the New York Times Expedition of 1886, as a result of which parts of Malaspina Glacier were briefly described by Lieut. Frederick Schwatka,[4] by H. W. Seton Karr,[5] by Prof. William Libbey, Jr.,[6] and the Topham Expedition of 1888, after which parts of the western Malaspina Glacier and its tributaries were described by H. W. Topham,[7] by George Broke [8] and by William Williams.[9] These explorations, however, were not made in that part of the Malaspina Glacier on which the authors of this book have made their recent observations.

It was not until Prof. I. C. Russell made his first expedition to Mt. St. Elias, in 1890, that the remarkable peculiarities of the Malaspina Glacier were made known;[10] and additional observations were made the following year in his second expedition.[11] On these two expeditions Russell crossed all the tributaries of the Malaspina between Yakutat Bay and the base of Mount St. Elias, and ascended several of them; he traversed the entire outer margin of the Malaspina from the Yahtse to Kwik River; he crossed the

[1] Tarr, R. S. and Martin, Lawrence, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 162–164; Tarr, R. S., Professional Paper 64, U. S. Geol. Survey, 1909, pp. 134–137.

[2] Dall, W. H., Ann. Rept., U. S. Coast Survey for 1875, Appendix 10, p. 157.

[3] Coast Pilot, Alaska, Pt. 1, Washington, 1883, p. 212.

[4] Schwatka, F., New York Times, Oct. 17, 1886; The Expedition of the New York Times, Century Magazine, Vol. XLI, 1891, pp. 865–872.

[5] Shores and Alps of Alaska, London, 1887, pp. 45–124; Alpine Regions of Alaska, Proc. Roy. Geog. Soc., Vol. IX, 1887, pp. 269–276.

[6] Some of the Geographic Features of Southeastern Alaska, Bull. Amer. Geog. Soc., Vol. XVIII, 1886, pp. 279–300.

[7] A Visit to the Glaciers of Alaska and Mt. St. Elias, Proc. Roy. Geog. Soc., Vol. XI, 1889, pp. 424–433; An Expedition to Mt. St. Elias, Alpine Journal, Vol. XIV, 1889, pp. 345–371.

[8] With Stock and Sack in Alaska, London, 1891, pp. 37–121.

[9] Climbing Mt. St. Elias, Scribner's Magazine, Vol. V, 1889, p. 387–403.

[10] Russell, I. C., An Expedition to Mount St. Elias, Alaska, Nat. Geog. Mag., Vol. III., 1891, pp. 53–203; Twelfth Ann. Rept., U. S. Geol. Survey, Pt. 1, 1891, pp. 59–61.

[11] Russell, I. C., Second Expedition to Mount St. Elias, Thirteenth Ann. Rept., U. S. Geol. Survey, Pt. 2, 1892, pp. 1–91.

glacier from the Yahtse to the base of Mt. St. Elias; and he made several excursions out upon the ice and to the bordering mountain side. From this comprehensive study he has given us an account of this remarkable ice plateau which has not since been improved upon; he made a general map (Pl. XI) which is the basis for all others; and he showed clearly that the Malaspina represented a hitherto unrecognized type of glacier to which he gave the name *piedmont* glacier.

In 1897 two expeditions started from Yakutat Bay with the object of ascending Mount St. Elias. Both crossed the Malaspina Glacier to the mountain base, using the glacier as a highway of travel as Russell had done. One of these, led by H. C. Bryant, was forced to abandon the attempt to reach the summit of the mountain, and unfortunately no description of the results obtained has been published. The other, led by Prince Luigi, Duke of the Abruzzi, succeeded in ascending to the summit of St. Elias, and in the report of the expedition[1] there is some description of the glacial phenomena, and some excellent photographs by Vittorio Sella are published.

Our own observations of Malaspina Glacier are principally upon the eastern edge where we traversed portions of its surface and some of its tributaries in 1905 and 1906, examining more distant parts of the glacier with field glasses in these years and in 1909, 1910, 1911 and 1913 and seeing parts of the western edge from the ocean in 1904 and 1910. For the sake of completeness, however, we shall review the conditions in the whole glacier, dealing with the western part of Malaspina briefly, except where we present fresh facts and discuss new interpretations of conditions there and more fully with the eastern portions where personal observations enable us to write with greater confidence.

Guyot, Tyndall, Libbey, Agassiz and Seward Glaciers

The western tributaries of the Malaspina (shown in Fig. 2 and Pl. XI) have been observed by the authors of this book only from a distance. No recent advances of these glaciers, similar to that of the eastern tributaries of the Malaspina described in this chapter, are known to have taken place. As the western glaciers are liable to future advances, similar to that of Marvine Glacier, it has been deemed proper to describe their history so far as it is known, especially as we are able to call attention to evidence of older advances, not previously summarized.

Guyot Glacier seems to have been severely crevassed from rapid normal movement throughout recent times and there are no known changes aside from the possible great advance above referred to.

Tyndall Glacier was apparently rather inactive in 1886–88, more active and crevassed in 1891, and perhaps not as active in 1897. The map published by the New York Times Expedition[2] and the Topham Expedition[3] indicate that Tyndall Glacier was not impassably crevassed in 1886 and 1888, for the routes of these parties go up near the middle of the glacier. A photograph from the Chaix Hills, taken by Russell in 1891, reproduced here as Pl. XII, A, shows such severely crevassed ice along the route followed three years before by the Topham party that a renewal of activity between 1888 and 1891 is thought

[1] Filippo de Filippi, The Ascent of Mount St. Elias by H. R. H. Prince Luigi Amedeo di Savoia, Duke of the Abruzzi, London, 1900.

[2] Seton Karr, H. W., Shores and Alps of Alaska, London, 1887, map on p. 87.

[3] See Broke, George, With Sack and Stock in Alaska, London, 1891, map facing p. 61.

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possible. Bryant's photograph of Tyndall Glacier from an adjacent peak in Chaix Hills in 1897 seems to show much less crevassing.

A rumor in Alaska in 1909 that a long narrow bay extending many miles back into the

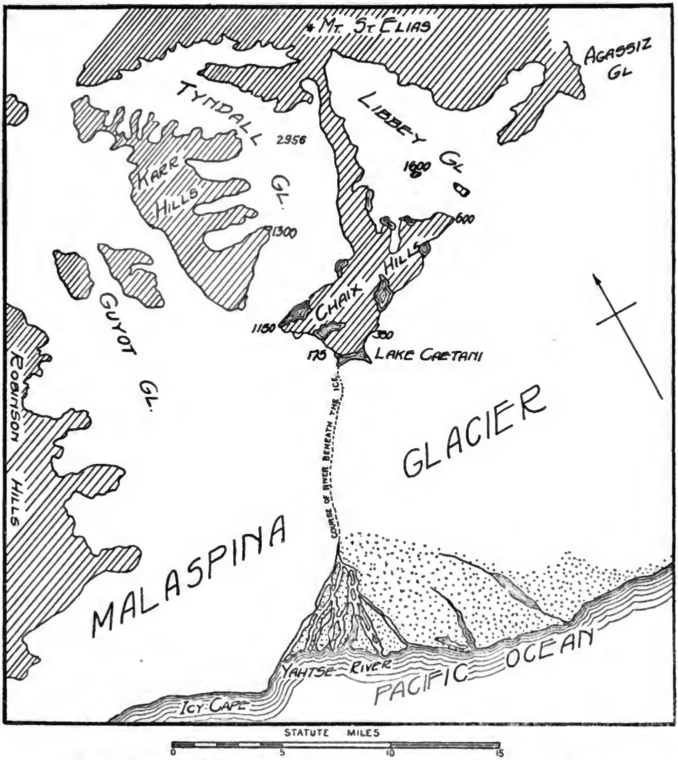


FIG. 2. TOPHAM'S MAP OF THE WESTERN PORTION OF MALASPINA GLACIER IN 1888.

western part of Malaspina had recently formed, suggested that Tyndall and Guyot Glaciers had either retreated very rapidly by the floating away of large ice masses or had advanced notably so that a new bay was formed along one side of the lobe. This was

investigated as far as possible in 1910 and a prospector, Mr. Joseph Pelagrini of Yakutat, who had recently been to the west edge of Malaspina Glacier, informed the junior author that a small cove does exist in the ice front near the Robinson Hills but that it is much smaller and shallower than has been commonly reported; that it is not on the site of the old Ice Bay but west of Guyot Glacier; and that it was not there on an earlier occasion when he visited this coast. He thought it had been formed by the floating out of an ice block.

In May, 1911, Mr. C. G. Quillian, commanding the United States Coast and Geodetic Survey steamer *McArthur* went within a mile of the western edge of the Malaspina Glacier. He examined the new Icy Bay at its western margin.[1] It is in such a position as to indicate that the Guyot lobe of Malaspina Glacier has receded several miles. The latest Coast Survey map [2] indicates that this recession of Guyot Glacier is about 9 miles, but Captain Quillian's estimate of the distance is only 2½ to 3½ miles. The discharging face of the glacier within the bay was 200 to 250 feet high in 1911. The recession which formed this bay surely took place later than October 4, 1904, when the junior author saw Icy Cape from a steamer close inshore, and before 1909.

Agassiz Glacier and its tributary, the Libbey Glacier, are normally somewhat crevassed in their mountain valleys, Libbey less than Agassiz, but their outer portions and extension into the piedmont Malaspina plateau were stagnant and more or less covered with ablation moraine when seen by Bryant in 1897 (Pl. XII, B), and by Russell in 1891. In view of this inactivity the advance of the border of Agassiz lobe on the edge of Chaix Hills in 1886 is worthy of notice. Schwatka states [3] that "at one point of the ice foot the glacier had shoved down into the timber, crushing into pulp and splinters huge trees five and six feet in diameter, as a child would sweep together his pile of jack-straws with his hand." Seton Karr [4] locates this point of advance on February 23 and 24, 1886, on an "island" south of the end of Chaix Hills, one of their camp sites, "This island was bordered on one side by the glacier, which was gradually advancing over it, crushing up the tall pines, rending them into matchwood, and heaping one over the other—a scene of gradual destruction by a resistless force. The onset of the glacier was overriding and burying the patch of wood." . . . "Close by, over the river and looming through the smoke, hung frowning cliffs of ice, the flank of the glacier-face which was burying our island; while, as if to add an additional horror to the scene, a tree crashed down at that moment, overborne by the weight of the advancing glacier." It is not explicitly stated by Schwatka and Seton Karr but it seems clear that this was an advance of Agassiz rather than Tyndall Glacier because of the ease with which the latter was traversed by the party during the next few days.

Topham, however, describes the continuation of this advance specifically in connection with his journey up the east side of Chaix Hills on the border of Agassiz Lake, in 1888.[5] This advance seems to have continued for at least two years, although it must have been nearly over in 1888 when they were able to travel along the edge of the ice. He

[1] Personal communication.
[2] Chart 8002, March, 1912.
[3] Century, Vol. XLI, 1891, p. 869.
[4] Op. cit., pp. 91, 96.
[5] Topham, W. H., Proc. Roy. Geog. Soc., Vol. XI, 1889, p. 428.

says "the Malaspina Glacier has shrunk away from the hills, and has left a moraine along their sides. Nevertheless, at one place, at an angle formed by a spur of the hills, the glacier is pushing up against the side of the hills and is crushing down the scrub trees and beautiful flowers. So fast is it doing this, that branches of alder, partially covered with stones and quite alive, are peeping forth from under the débris and protesting against the encroachment of the ice. This débris consists for the most part, not of stones brought along upon the surface of the ice, but of an old moraine, which is being overwhelmed and crushed. I believe that this ice is sliding and swelling over the older ice below so that it can have little or no effect upon the Malaspina Glacier taken as a whole. There are no signs along the edge of the latter down by the mouth of the Yahtsé-tah, that it is either advancing or receding There are no piles of stones left behind to indicate its retreat, and no trees crushed down to show its advances."

The upper parts of Agassiz and Seward Glaciers, apparently normally crevassed because of more rapid movement, have been vividly described by Russell and by Abruzzi. Seward Glacier (Pl. XI), perhaps the largest valley glacier in North America, with its known length of 25 miles and unknown extension of perhaps 25 miles more, its average width of 3 to 6 miles and its scores of great tributaries, is of especial interest as it is the largest tributary of the Malaspina.

In 1890 Russell described the Seward Glacier within its mountain valley rather fully,[1] making measurements of its rate of movement which seemed to be about 20 feet a day in the central part of the ice stream.

Seven years later the Abruzzi expedition traversed parts of Seward Glacier, taking many splendid photographs which will be of value for determining future conditions in Seward Glacier, and noting especially[2] a greater amount of crevassing than in 1890. The motion, however, was apparently slower in 1897 than the rate measured by Russell for Filippi calls attention to the fact that "Mr. Russell and his fellow-explorer, Mr. Kerr, both relate how seracs frequently crashed down with such force as to shake the ice under their feet, and they add that almost incessant reports and rumblings were produced by the rolling and shattering of the fallen rocks. Nothing of the kind was observed by ourselves during the days we spent on and about the Seward. The glacier was always perfectly quiet; only now and then a solitary stone would come down, or a fragment of sérac would drop into a crevasse with a dull thud." A possible interpretation of the greater crevassing but slower movement in 1897 would involve a period of more rapid movement just beginning in 1890 (Russell unfortunately did not go back this way in 1891), but which had ceased before 1897, although the glacier was still severely crevassed the latter year as a result of it.

In 1911 Seward Glacier was easily crossed near Pinnacle Pass by a Boundary Survey party. Their photographs indicate that the crevassing was about the same as in 1897.

In 1891 Russell observed near Point Manby evidence of a former advance of the Malaspina border due to earlier renewal of activity of either Seward or Marvine lobe.[3] "A recent advance of the glacier had cut scores of spruce trees short off and piled them in confused heaps. After this advance the ice retreated, leaving the surface strewn with an irregular sheet of boulders and stones. . . . The glacier during its advance

[1] Nat. Geog. Mag., Vol. III, 1891, pp. 177–180.

[2] The Ascent of Mt. St. Elias, London, 1900, p. 111.

[3] 13th Ann. Rept., U. S. Geol. Survey, 1894, p. 63.

ploughed up a ridge of blue clay in front of it . . . thickly charged with sea shells of living species."

In neither 1905, 1906, nor 1909, did we make any observations on Seward Glacier or the Seward lobe of the Malaspina Glacier excepting to look out upon it from elevated points near the eastern margin of the Malaspina. Even such views, distant though they were, were sufficient to show that the Seward lobe had not in any one of these years felt the impulse of a forward thrust similar to that which swept down the Marvine lobe in 1906. From Blossom Island, in 1906, an area of crevassing was seen near the point where Seward Glacier emerges from its mountain valley which was thought to possibly represent the beginning of an advance, for one of the 1906 party, Mr. Alexander, who was on the Abruzzi Expedition in 1897, was sure that the crevassing was much greater the latter year in parts of the glacier sledged over by him nine years before. So far as could be seen in 1909 no great advance had taken place in the interval. The surface of the Seward lobe appeared to be as smooth as formerly, though near the mountains there is a very irregular ice surface which is probably normal, for photographs by the Duke of Abruzzi's party in 1897 also show crevassed ice in approximately this position.

It may be pointed out that it is quite possible for a large glacier, with many tributaries, to have more than one advance, with different degrees of intensity, as an outcome of earthquake shaking. The advance of one tributary, or of several combined, might cause slight advance, and later another greater advance might result from the thrust derived from a larger glacier or several tributaries. There is no proof that this has been the case in the Seward Glacier and it is merely stated as an interesting possibility, and in order to make it clear that a slight thrust may have occurred in 1906 and yet, after an interval of years, a still greater advance might begin.

Marvine Glacier

Stagnant in 1890. Russell describes Marvine Glacier within its mountain valley as partly covered with "interminable fields of angular débris,"[1] with areas of "hard blue ice," some medial moraines, "broad areas covered with sand cones and glacier tables," but apparently with no amount of crevassing such as would interfere with rapid packing over its surface.

The Marvine lobe of Malaspina Glacier seems to have also been inactive in 1890 and possibly at least as long before as 1874, as is suggested by the observation of the botanist Coville.[2] In 1899 with a small party from the Harriman Expedition he spent several days on the eastern border of Malaspina Glacier about half way from Disenchantment Bay to Point Manby. Here the ice at the border of the Marvine lobe was covered with a thin layer of soil which supported vegetation. There were no spruces growing on the ice but the oldest alder bushes were perhaps 8 feet high and 25 years old.

Little Changed in 1905. On August 22nd, 1905, the junior author climbed up from Hayden Glacier and looked out upon the Marvine and Malaspina Glaciers, with field glasses, from an elevation of 1585 feet on the Floral Hills. At that time the valley portion of Marvine Glacier and a broad belt along the eastern margin of the Marvine lobe

[1] Nat. Geog. Mag., Vol. III, 1891, pp. 122–123 and Pl. 17.

[2] Coville, F. V., Personal communication.

of Malaspina Glacier were covered with ablation moraine. This was recorded by notes and in a sketch map and photographs (Pl. XIII, A). Dirty, somewhat crevassed ice was observed along the border of the Hitchcock Hills toward the point of emergence of Seward Glacier, in Malaspina Glacier down to what was then taken to be Kame Stream, and thence eastward parallel to the glacier margin. Outside this moraine belt was clear ice with the morainic scrolls previously alluded to. There was no lake visible near Blossom Island. The Marvine Glacier, however, was apparently not crevassed, in its valley, and seems to have been almost exactly as when traversed by Russell in 1890. The whole eastern Malaspina Glacier, although not seen from nearer than this, seems to have been in the fall of 1905 essentially as it had been when crossed by the expeditions of 1890, 1891, and 1897, except in the portion near the east border where crevassing had then commenced. Near the débouchure of Marvine and Hayden Glaciers the moraine-veneered ice was broken into flat-topped tables or seracs, but practically none of the débris had as yet disappeared into the crevasses. Its condition was not unlike that of lower Lucia Glacier in 1909. In view of the remarkable extension and continuation of crevassing observed by the senior author in 1906, this beginning of crevassing in 1905 is of much interest,[1] especially as this wave of advance which produced the crevassing seems to have appeared in the piedmont bulb before it affected the valley glacier. This area of crevassing is similar to the much smaller one seen in Atrevida Glacier the same year. Our impression was that even the crevassed area could have been crossed in August, 1905; but in 1906 this part of the Malaspina Glacier was impassable.

Advance of Marvine in 1906. The expedition of 1906 had for its object the crossing of Malaspina Glacier from east to west and the further study of this piedmont ice mass; but by a sudden and most unexpected change in the condition of the eastern or Marvine lobe this plan was thwarted. The nature of this change, as observed along the eastern margin of the Malaspina Glacier from Point Manby to Blossom Island, has been fully described [2] elsewhere, and will only be summarized here.

Where Marvine Glacier emerges from its mountain valley above Blossom Island the progress of the expedition of 1906 was absolutely barred by a sea of crevasses and bristling pinnacles (Pl. XIV), the entire glacier surface from side to side, for a width of fully 4 miles, being broken as only rapidly-moving glaciers are, and resembling an ice fall in the intensity of the breaking. Yet it was at this very point that Russell had easily crossed in 1890, carrying his entire outfit toward Mount St. Elias. From this point down to the sea the surface of the Marvine lobe was broken into an impassable condition, and the margin of the glacier was also continuously broken except where joined by the Hayden Glacier, by which the area of crevassing was pushed out from the mountains. We could not get far out upon the glacier at any point along the margin, for everywhere the way was barred by an impassable network of crevasses. This condition of breakage extended even to the shores of Yakutat Bay, and across the routes followed by Russell on his retreat in 1891 and by the Bryant and Abruzzi expeditions in 1897. The area

[1] The fact of this beginning of crevassing in 1905 was not seen by the senior author, who did not cross Atrevida, Lucia, or Hayden Glacier in 1905. It was not known to him when he wrote the several papers on the 1906 conditions in which he has assumed that the crevassing began between August, 1905 and June, 1906. We did not see the area together in either 1905 or 1906 and the recorded observation of crevassing in 1905 has only been noticed since our 1909 expedition.

[2] Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, pp. 83–88; Tarr, R. S., Bull. Geol. Soc. America, Vol. 18, 1907, pp. 273–277.

of crevassed ice, some 4 miles wide where Marvine Glacier emerges, broadened toward the sea (Fig. 3), being about 5 miles wide opposite Blossom Island, and on the shores of Yakutat Bay extending from the Kwik River at least as far as Point Manby. For a distance of over 15 miles, from the mountains to the seaward face of the Malaspina, the ice was crevassed over a width increasing from about 4 miles near the mountains to 12 or 13 miles near the sea.

Near Blossom Island the senior author found this lobe of the glacier very different from its condition when Russell and others walked and sledged over it. It was im-

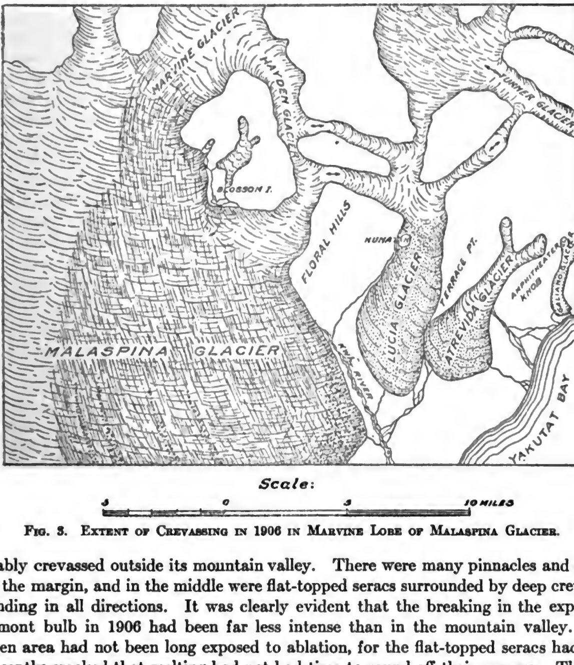


FIG. 3. EXTENT OF CREVASSING IN 1906 IN MARVINE LOBE OF MALASPINA GLACIER.

passably crevassed outside its mountain valley. There were many pinnacles and arêtes near the margin, and in the middle were flat-topped seracs surrounded by deep crevasses extending in all directions. It was clearly evident that the breaking in the expanded piedmont bulb in 1906 had been far less intense than in the mountain valley. The broken area had not been long exposed to ablation, for the flat-topped seracs had been so recently cracked that melting had not had time to round off their corners. The fact that not only the snow-covered, but even the bare serac-tops were not notably affected by melting proved that most, if not all, of the breaking had taken place after the period of ablation of 1905 had ceased.

Sledging over this surface, which Russell and others had crossed easily, would have

been impossible in 1906 and even travel on foot could be undertaken only by the most difficult ice work.

We traversed the eastern margin of Malaspina Glacier from the point of emergence of Marvine Glacier to the mouth of the Kwik River, and along this entire distance, excepting only where dominated by Hayden Glacier, the ice was greatly broken. Prior to 1906 this land margin had been covered with ablation moraine bearing a forest of cottonwood and alder and some spruce along the margin of the lower Kwik valley, and this margin was so moderately sloping that it could easily be ascended at any point; in 1906, however, the margin was for the most part changed to a steep and broken cliff down which the ablation moraine was constantly sliding, while in the forest-covered area trees and bushes were overturned and were falling into crevasses or crashing down the newly steepened ice face (Pl. XV, A). The cliff was crevassed, often, into impassable condition, and the broken ice blocks protruded through the ablation moraine, giving the appearance of a frost-riven rock cliff, heightened by the earthy stain that the angular ice blocks bore, and by the trees that still stood among them. The ablation moraine, which had covered the ice for a depth of from 2 to 15 feet, and had so protected it from melting that sufficient stability was secured for the growth of trees to maturity, was now sliding into crevasses and down the ice cliff, often in mud flows. Abundant water (Pl. XVI, A) was now supplied by the melting of the ice, newly-exposed to the air and rains, and small and large streams, heavily charged with morainic débris were issuing from all portions of the ice margin, where hitherto only trickling streams of clear water slowly oozed from the moraine-covered glacier.

That the ice was being broken and crowded forward during July and August, 1906 was evident as we passed along its base; and that the entire breakage was a result of changes during that year was indicated by the fact that among all the trees found overturned, not one was seen that had not fully developed foliage (Pl. XV, B). This was the last season for many trees, for in August some were dead and others had yellow leaves, as if dying. This evidence is in harmony with several other facts which indicate that this advance was confined mainly to 1906. These facts are as follows: (1) Our observations of 1905, though not very full, indicate that the advance had barely begun; (2) other glaciers in the region underwent similar absolute transformation in the brief period of nine months; (3) the slight amount of ablation on the broken surface of the Marvine lobe, already mentioned; (4) the closing of the subglacial tunnel by which the waters of the Blossom Island region had formerly escaped, which occurred between the autumn of 1905 and the middle of July, 1906, and which gave rise to the formation of a temporary expansion of the lake which lies in the Blossom Island depression.[1]

Thus a long period of stagnation of sufficient duration to permit the growth of a forest with trees a half century old, and in other portions of the glacier a period of such moderate motion as to permit the development of a smooth ice surface over which travel was easy in all directions, was abruptly ended by such a sudden and rapid forward motion that even the most stagnant portions of the ice margin were broken. Whether the ice was thickened by the advance could not be determined, for we had no exact basis for measurement, but, in view of the fact that other glaciers which advanced were thickened, it is probable that this one was also. There was certainly some advance along the margin, for only by this could the ice cliff be steepened and broken, and ice blocks caused to

[1] Tarr, R. S., Professional Paper, U. S. Geol. Survey, 1909, p. 88.

fall from its front; and, moreover, there were places where it had encroached on the fringing forest; but there was no basis upon which we could determine the amount of forward motion. It is probable, however, that the amount of advance and thickening was not nearly so great here as in other glaciers, for there was a broad area of piedmont ice in which the energy of the forward thrust could be dissipated, that it is hardly to be expected that extensive thickening and advance could also be caused. It is remarkable that so vast an area of piedmont ice should have been affected (approximately 300 square miles) and this could only have resulted from a thrust of great vigor from the Marvine Glacier on its emergence from the mountain valley.

Again Inactive in 1909. In the report on the expeditions of 1905 and 1906,[1] it was predicted that the new cycle of advance of Marvine lobe would probably end in a treeless and moraine-free glacier margin after another season's melting had destroyed the forest and removed much of the moraine. Speaking of the advancing glaciers in general it was inferred that in each case stagnation would soon follow the dying out of the wave of advance, after which ablation moraine would again cover the glacier margins. The latter prediction was fully borne out by our studies of 1909, but the former was not.

By the time we approached the margin of the Malaspina Glacier in 1909 we had already satisfied ourselves as to the main elements in the problems of the advancing glaciers, and felt little need of a comparative study of the Malaspina margin, which we had at first thought of undertaking. To have done so meant the entire abandonment of our plan of visiting the Prince William Sound region, for the difficulties of travel up Kwik valley are so great that several weeks would be required, and the route through Floral Pass was closed to us by the recent advance and resulting broken surface of Lucia Glacier. We, therefore, undertook nothing further than an examination of the Marvine lobe from elevated points at a distance, from which, however, we could clearly see its condition even with the naked eye, and satisfied ourselves that it showed no new features of importance that were not exhibited by other advancing glaciers of the 1906 group. In 1910 the ice jam in Yakutat Bay made it impossible to land on the west side of the bay and examination of Marvine lobe with field glasses added no significant facts to those observed the previous year. In 1911 Marvine Glacier was once more so little crevassed that it was easily crossed in its mountain valley by a Boundary Survey party. In September, 1913, the Marvine lobe of Malaspina Glacier was examined from a distance with field glasses by the junior author. It seemed to be entirely stagnant and inactive, the margin being again covered with ablation moraine.

All these observations clearly indicate that the advance of 1906 was short-lived, and that by July, 1909, it had completely ceased. Moreover, ablation had so healed the broken surface and margin that we were convinced that the advance ended soon after August, 1906. In the clear ice area of the Marvine lobe back of the margin, the ice surface looked much as it did in 1905; similar swirls of moraine were seen and the surface looked quite smooth. Doubtless a nearer view would have shown that the surface was far rougher than in 1890 and 1897 but it was certainly much less irregular than in 1906, and it appeared probable that one could travel over the glacier surface with no great difficulty.

The eastern margin of the Malaspina was also noticeably healed, as compared with the 1906 condition, though still rough, broken and steep. The jagged marginal cliff had disappeared, though steep ice cliffs still appeared in places, and while much of the

[1] Tarr, R. S., Professional Paper 64, U. S. Geol. Survey, 1909, pp. 87 and 94.

forest that grew in the ablation moraine had gone, much alder and cottonwood still remained. The destruction of neither the moraine nor the forest was complete, but the margin presented a far different appearance from that of 1905. Then the forest-bearing portion of the glacier margin was a moderately sloping embankment covered with forest verdure; in 1909 there were wooded patches interspersed with areas of barren moraine, and even with ice cliffs visible from a distance. In view of the rapidity with which the moraine was being removed from the ice in 1906 and the forest being destroyed, the fact that no more destruction had been accomplished makes it difficult to believe that the period of advance and breakage lasted even into the summer of 1907. With cessation of motion, ablation in the low marginal areas would quickly bring about a sufficient degree of soil stability for vegetation already growing to maintain itself with little further destruction by undermining. However, that ablation was still proceeding rapidly in those parts of the glacier not protected by a deep ablation moraine was indicated by the volume of the Kwik River, which appeared to be even larger in 1909 than in 1906. That such a speedy ending of the 1905–6 advance, and the consequent rapid healing of the broken surface by ablation was not peculiar to the Marvine lobe of the Malaspina Glacier is made clear in later pages where other advancing glaciers are considered.

Hayden Glacier [1]

This glacier has already been mentioned as one of the tributaries of the Malaspina Glacier, though contributing little ice to it. Heading on the western slopes of Mount Cook the Hayden Glacier flows as a broad, moderately-sloping valley glacier (Pl. XIII, B), with a width of 3 or 4 miles, and, on emergence from its mountain valley, expands to almost double that width, coalescing with the Marvine lobe of the Malaspina. A moderate recession would disconnect it from the Malaspina, giving it independent existence, such as Atrevida and Lucia Glaciers, once tributary to the Malaspina, now have. It is somewhat crevassed in places, and below the snow line, just west of Floral Pass, is bordered by lateral moraines, while on the expanded portion, outside the mountain, are several medial moraines and morainic swirls of interesting pattern. The amount of moraine on the surface increases toward the Marvine lobe, giving rise to many morainic ridges; but nowhere except along the margins is there a continuous cover of ablation moraine.

This glacier was partly crossed by the junior author in 1905 and observed and photographed from the western slopes of Floral Hills; and it was crossed and recrossed several times, and along several routes by the senior author in 1906. Travel across its surface in any direction was not difficult and it was evident that there had been no noteworthy change in condition since Russell crossed it in 1890. We did not get as far west as the Hayden Glacier in 1909 for the broken Lucia Glacier stood in the way, but we saw portions of its surface with field glasses in 1909 and 1910 and were convinced that it had not undergone any noteworthy change in the interval between 1906 and 1910. This negative result is interesting in view of the fact that Marvine Glacier, the next large glacier toward the northwest, and Lucia and Atrevida Glaciers, its large neighbors toward the southeast, have both felt the impulse of the earthquake advance. One may anticipate that its turn will soon come.

[1] For a fuller description, see Tarr, R. S., Professional Paper 64, U. S. Geol. Survey, 1909, pp. 81–83.

CHAPTER IV

LUCIA AND ATREVIDA GLACIERS

Lucia Glacier

General Description. Lucia Glacier, which is between fifteen and twenty miles long and about two miles wide, is a valley glacier expanding in a piedmont bulb outside the mountain front, and in its lower portion covered with a broad waste of ablation moraine. Its sources are unknown, but in the main, if not entirely, are on the slopes of Mount Cook. That there is no large tributary heading farther back in the St. Elias Range is indicated by the morainic débris, which includes none of the crystalline rock of which these more remote mountains are composed. While the exact sources are unknown, it is certain that the glacier receives many tributaries, and some good-sized ones from both sides are visible from the lower portion of the glacier.

Opposite Floral Pass, where the Lucia Glacier is enclosed on both sides by mountain valley walls, the width of the glacier is 2 or 3 miles, but it expands beyond this point because the valley becomes wider and further down because the valley wall on the eastern side terminates. The width of the expanded piedmont bulb is 4 or 5 miles at the broadest part, the glacier being prevented from further expansion by the presence of the Floral Hills on the west and by the competing piedmont bulb of Atrevida Glacier on the east. The outer portion of the Lucia extends southward to the Kwik valley, forming one wall of that valley, while the Malaspina border, a mile or two away, forms the opposite wall. The Lucia and Atrevida piedmont bulbs coalesce and form a small piedmont glacier, similar to the great Malaspina, but supplied with ice from only two tributaries (Map 2, in pocket).

Basis for Study. This glacier, as well as the Atrevida, was crossed by Professor Russell and named by him in 1890.[1] He first used this name, however, for Seward Glacier.[2]

It was next visited by the Geological Survey expedition of 1905, during which its lower portion was examined from a neighboring mountain and was crossed by the junior author and studied late in August. In the Geological Survey expedition of 1906 access to the glacier from the east was cut off by the advance and breaking of the Atrevida Glacier, but it was reached and crossed from the west by the senior author.[3] The National Geographic Society's expedition of 1909 traveled to the eastern margin of the Lucia Glacier, but further exploration was checked by the advance of the glacier. We were, however, able to examine it from Ampitheatre Knob and from an elevation on Terrace Point, just above its eastern margin, from which we could see all parts of the glacier except the upper portion far back in the mountains. The description of the 1909

[1] Nat. Geog. Mag., Vol. III, 1891, pp. 92, 105–108.

[2] Twelfth Ann. Rept., U. S. Geol. Survey, 1891, p. 60.

[3] For a description of this glacier in 1905 and 1906, from which portions of the following description are abstracted, see Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1906, pp. 75–80.

condition is based upon these observations. The glacier was seen only from the east side of Yakutat Bay in 1910.

Description by Russell. Lucia Glacier was inactive in 1890, for Russell described it during the first part of July, 1890, as "covered from side with angular masses of sandstone and shale . . . while further up the valley the débris on the surface of the ice disappeared, and all above was a winter landscape. The brown desolate débris fields . . . extended far southward, and covered the expanded ice foot in which the glacier terminates. Most curious of all was the fact that the moraines on the lower border of the glacier were concealed from view by a dense covering of vegetation, and in places were covered with forests of spruce trees." He also described the nunatak on the west side of the glacier and the outwash plain at its base.

Continued Inactivity in 1905 and 1906. Lucia Glacier seems not to have changed significantly between 1890 and 1905; and between 1905 and 1906, there was no notable difference, except a possible increase in crevassing up the valley to which reference will be made later. In 1905 it was impossible to clearly detect the boundary between Lucia and Atrevida ice tongues, the two lobes of this piedmont glacier; though for a part of the distance a line of alder growth, forming an oasis in a broad waste of ablation moraine, marked the site of the interlobate area of greatest stagnation. Above this the two glaciers were separated by the gravels of Terrace Point, and near the outer border of the piedmont glacier a stream emerged and formed a valley, apparently in line with the boundary between the two lobes; but for some distance above this valley the two lobes so coalesced that the boundary between them was indistinguishable. In 1906, however, the advance and crevassing of Atrevida Glacier, which pushed westward into the still stagnant Lucia lobe and destroyed the interlobate strip of alder, brought out sharply the area occupied by each lobe (Pls. XXVI and XXVII). The Atrevida lobe was everywhere broken, with clear ice showing along the boundary between the two lobes, while the Lucia was unchanged—a broad, undulating waste of ablation moraine which, from a distance, bore little resemblance to a glacier.

One of the most interesting features of the Lucia Glacier was its ablation moraine; it seemed stagnant near its terminus and wasting in its valley and was the best illustration of this type in the Yakutat Bay region. For over 7 miles, from its terminus to a point well up the mountain valley, the glacier surface was obscured by ablation moraine (Pl. XVII, A), and throughout fully nine-tenths of the area no ice could be seen from a distance. The ablation moraine extended well up the valley portion of the glacier, to a point at least a mile above Floral Pass, becoming rapidly thinner, with more ice showing up the valley, and finally giving place to bands of medial and lateral moraine. Opposite and below Terrace Point the moraine increased in proportion and ice appeared only in small bands, or in depressions holding lakelets, at least a score of which dotted the moraine-covered portion of the glacier. The surface of this desert waste of ablation moraine was nowhere notably crevassed, and travel over the surface was possible in all directions, the only difficulty being the extreme irregularity of the surface due to differential melting, and the occasional slippery slopes where the moraine veneer was thin. There were numerous moulins of large size, and morainic débris was being carried into these by the drainage. This was the only escape for the concentrated morainic débris, for there were no crevasses of any considerable depth in this slowly moving glacier.

For two miles or more the outer margin of the Lucia lobe was clothed with vegetation. On the inner side of this forest-covered part of the ablation moraine scattered individuals and clusters of alder were growing in the more stable portions of the moraine, but the amount of vegetation rapidly increased toward the outer margin of the glacier, and in a short distance the entire surface was covered and hidden by alder thickets. Farther out individual cottonwoods rose above the alder, and at the extreme margin a forest of cottonwood and spruce occupied the ablation moraine. The forest covered more than a mile of the outer glacier margin and in it no ice was visible from a distance. The extreme outer margin consisted of a low, wooded, gently-sloping embankment which rose above the gravels of the Kwik valley, and it was exceedingly difficult in places to tell where the glacier ended. Without examining it carefully, one unfamiliar with the phenomenon of forest-covered glaciers would not suspect that this wooded slope was a glacier end. One noticeable feature, however, was the large number of small, trickling streams of clear, cold water that emerged from the embankment, and at one or two points larger streams, bearing sediment.

In the ablation moraine were many interesting details, most of which have no direct bearing on the glacier as a whole but are phenomena of melting. There were, however, several noteworthy features. The most prominent of these was the extreme irregularity of the surface due to differential melting. In crossing the ice surface one was constantly climbing over knolls and ridges and descending into valleys, some roughly circular, others linear or crescentic, and the difference in elevation from valley bottom to ridge crest was often fully 100 feet. The elevations occurred where the drift was thick, the depressions where it was thin, and it was evident that the relative positions of these higher and lower portions of the moraine surface were constantly changing as the thickness of the drift cover changed through downsliding. During the ablation of the under ice, steep slopes are developed down which the débris slides to the depressions, leaving nearly or quite bare ice slopes, while the débris accumulates in the depressions and in time doubtless accumulates to sufficient depth to retard their further reduction. Then the depressions are gradually transformed to elevations, and the former elevations, denuded of their drift cover, are lowered to valleys. Where the drainage escapes into moulins there are conditions favoring more permanent valleys. It is this constant shifting of débris that prevents the growth of vegetation on these thinner portions of the ablation moraine, and the struggle that plants are making to find a foothold here is often illustrated near the margin of the plant-covered portion of the ablation moraine, where annual plants and one or two year old alder and willow bushes are found overturned, or dying where they stand by the removal of the soil from around their roots.

A second noteworthy feature of the ablation moraine was the fact that the rock fragments were all angular, so far as we were able to observe. They were frost-riven fragments and avalanche falls derived from the steepened upper valley walls and in part, doubtless, supplied with the snow itself as it slid down into the reservoirs. In the outer part of the piedmont bulb the ablation moraine consisted of a confused maze of hillocks and depressions, with no definite system, and with a topography reminding one of some areas of sand dunes where ridges are absent; but farther up the glacier, where the moraine was thinner, the ablation moraine presented a series of crescentic ridges, where longitudinal ridges, perhaps medial moraines, swung across the glacier instead of down it. The cause for this peculiar arrangement of the morainic ridges is not clear, but is evidently

in some way connected with the flowage of ice under the conditions of expansion into a piedmont bulb of semi-stagnant or stagnant condition.

A final feature of note was a narrow depression between morainic ridges which in 1905 and 1906 ran eastward from the base of Lucia Nunatak, for about a half mile. It was much smaller than similar ice-flats seen on Variegated Glacier in 1905 and on Atrevida Glacier in 1909. The development of crescentic ridges and interior flats seems to be a phenomenon common to piedmont ice bulbs, and other instances will be pointed out in the descriptions of other glaciers. Their significance will be discussed in a later chapter after the various instances have been described.

On the western side of Lucia Glacier, almost at the mouth of Floral Pass, a low hill rises between 700 and 750 feet (Pl. XVI, B). Russell described it in 1890[1] as "a huge rounded dome of sandstone rising boldly out of the ice. This is similar to the 'nunataks' of the Greenland ice fields, and was covered by ice when the glaciation was more intense than at present. On the northern side of the island the ice is forced high up on its flanks and is deeply covered with moraines; but on the southwestern side its base is low and skirted by a sandplain deposited in a valley formerly occupied by a lake. The melting of the glacier has in fact progressed so far that the dome of rock is free from ice on its southern side, and is connected with the border of the valley toward the west by the sandplain. This plain is composed of gravel and sand deposited by streams which at times became dammed lower down and expanded into a lake. Sunken areas and holes over parts of the lake bottom show that it rests, in part at least, upon a bed of ice."

From its position this hill is useful as a register of changes in the glacier, and for that reason calls for detailed description. In 1905 and 1906 the eastern face of the nunattak was a steep cliff at whose base the glacier flowed in contact with the rock. The northern side, less steeply sloping, was also in contact with the ice, but the southern, or lee slope was free from ice, although the glacier spread westward just below it. On the western side a small ice tongue, or distributary of Lucia Glacier protruded part way down to Floral Pass with a slope of 12°, ending in a low, moraine-covered terminus. Thus the hill, which is about ¾ of a mile long and almost half as wide, had ice contact for about half its periphery, while ice was present at a short distance both to the west and south, being absent from less than a quarter of the periphery, toward the southwest. The ice on the northern slope rose more than two-thirds of the way to the top of the nunatak.

Fortunately the junior author made a careful section of the north side of Lucia nunatak in August, 1905. The ice rode up on this stoss side, at an angle of about 22°, to a height of 550 feet above the base where there was a rude moraine terrace, 100 yards or so in width, with an ice foundation still present (Fig 4). Above this were two earlier moraines 155 and 50 feet respectively below the top of the nunatak. The lower moraine was about 75 feet wide and the upper contained small lakelets. These 1905 relationships are shown in Pls. XVI, B and XVIII where the 1905 profile is contrasted with that of 1909.

The height to which the ice on the north side of the nunatak rose above the level of the main glacier, in 1905–6, made it seem unlikely that it could have been crowded up to that level by the then-existing thrust of the glacier. The natural inference from this condition is that some recent thrust had pushed the ice up here, and this inference is supported by the fact that for fully 100 feet above the ice the hill slope was barren of

[1] Nat. Geog. Mag., Vol. III, 1891, p. 106.

vegetation, although there was abundant plant growth above this zone. That part of the glacier which rose above the glacier level on the northern slope of the nunatak was completely covered with a thick coat of moraine by which ablation must necessarily be greatly retarded, so that the thrust which pushed the glacier up may well have occurred years ago, probably long before Russell's visit in 1890. By the advance of 1909 the ice condition around Lucia Nunatak has been greatly changed, as is stated below.

The western margin of the glacier, below Lucia Nunatak, was bordered by a large marginal stream which emerged from an ice tunnel at the southeastern corner of the nunatak and joined the Kwik River, receiving on the way, first a small tributary from the ice tongue west of the nunatak, then land streams from Floral Pass and other valleys in the Floral Hills, and small streams from the glacier. Lucia Stream was an impassable glacial torrent where it emerged from the ice tunnel, but it was swollen to nearly twice that size before it united with the Kwik. Although this stream flowed along the western margin of the glacier it was not everywhere in contact with it, for Lucia Glacier crowds against projecting spurs of the Floral Hills and the stream was at such places forced into rock-walled gorges cut across these spurs. Thus the marginal stream valley consists of stretches where the mountain forms one wall and the glacier the other, alternating with short sections enclosed on both sides by steeply-rising rock walls, against whose base the river in places flowed, forcing us, in our attempt to pass up the valley, out of the gorge upon the higher wooded slopes of the projecting spurs.

Conditions on the eastern margin of Lucia Glacier were quite different. Above Terrace Point the snow line was soon reached and the glacier was then in contact with the mountain, but at Terrace Point a broad valley existed, with the moraine-covered ice for one wall and the gravels of Terrace Point for the other wall. In this valley was a small stream, supplied by drainage from the land and from the ice, which disappeared beneath the ice, but which had earlier in the season been dammed to form a lake. Lacustrine deposits and stream gravels had accumulated here, and the process of formation of the gravels of Terrace Point was here being duplicated and illustrated in miniature. One or two small lakes existed farther down in this marginal valley in the depression between Atrevida and Lucia Glaciers, and below that came the area of coalescence between the crevassed Atrevida and the stagnant Lucia Glacier. Beyond this junction the two lobes flared apart and there was a valley, broadening rapidly westward to a V-shape, in which a small stream flowed over an alluvial fan, bordered on the west side by the low, wooded embankment of Lucia Glacier and on the east side by the similar slope of Atrevida Glacier. This stream evidently carried the drainage from the Terrace Point region plus such water as escaped from the eastern margin of the moraine-covered Lucia and the western margin of Atrevida Glacier. A large part of this water had passed beneath the stagnant interlobate area between the two glaciers. Evidently, therefore, the main drainage of Lucia Glacier was on the west and of the Atrevida on the east side, and the interlobe drainage was moderate in amount. This condition was greatly changed in 1909.

Lucia Glacier has at some former time been far more extensive than now. This is proved by the morainic deposits that cover Terrace Point, and by the moraine terraces that lie high up in Floral Pass and in other valleys in the eastern slopes of the Floral Hills. It has at some period been tributary to the Malaspina, and, while there is no proof of it, it is entirely possible that ice may even at present lie beneath the gravels of

the Kwik valley that now separate it from the Malaspina. These deposits are so thick that they would suffice to almost prevent further melting.

Advance of Lucia Glacier in 1909. So far as we know, Lucia Glacier was receding continuously between 1890 and 1909. In 1906 there was some evidence, though not very conclusive, of a coming change, and an advance was predicted on the basis of an apparent increase in crevassing up the valley and on the assumption that a longer glacier, like the Lucia, might reasonably be expected to advance later than the shorter ice tongues, like the adjacent Atrevida.[1]

When we saw Lucia Glacier in July 1909 it was advancing as the Marvine and others were doing in 1906, but had reached a less complete stage of advance than any of the 1906 glaciers. In view of the rapidity with which the advance passed down the glaciers, once it had started, as observed in 1906, it seems doubtful whether the crevassing observed in the upper Lucia Glacier in 1906 was in reality the forerunner of the 1909 advance. Had it been an indication of coming change one would expect that its effects would have been felt in the lower glacier by 1907. Still, our knowledge of the behavior of glaciers advancing under the impulse of snow supplied by earthquake shaking is too slight and too fragmentary to warrant definite exclusion of this suggestion of coming change. It may be true, for example, that only one or two of the tributaries had communicated the impulse of advance to the main glacier and that sufficient force had not yet been supplied to extend the crevassing farther down, while a year or two later the impulse from other branches, either more numerous or larger, gave sufficient impetus for a continuation of the thrust and for the extension of its effects down the glacier. This is an hypothesis at least worth retaining until the behavior of such advancing glaciers is better understood.

Be this as it may, Lucia Glacier, when seen in July, 1909, almost exactly three years after our last observation of it in 1906, was utterly transformed, and was rapidly changing under our very eyes. Between Terrace Point and Lucia Nunatak, where we had so easily crossed in 1905 and 1906, the undulating moraine-covered surface was rent by a series of gashes (Pl. XVIII) which rendered the glacier impassable. These gashes were profound, flaring crevasses of great length, spaced a few yards apart, nearly parallel, and extending in the direction of flow of the glacier. The glacier had the appearance of being newly-crevassed by differential strain due to motion down the valley and toward the center of the glacier there was some indication of faulting, with the uplift on one side of the crevasses. The direction of the crevasses varied both below and above Terrace Point, but in each case changes in direction seemed attributable to change in direction of flow of the glacier. For example, near Lucia Nunatak the crevasses pointed toward the valley center, as if the ice flow here were deflected by the nunatak; and in the piedmont bulb portion the crevasses fanned out as the ice itself does to form the piedmont expansion. Between these parallel linear gashes there were very few, in fact scarcely any cross breaks, but there were long, undulating strips of serac, still covered with ablation moraine, over which travel would be easy; and it seemed probable that once one reached this gashed area it would be a matter of no great difficulty to cross the central part of the glacier by a zigzag course, threading one's way around the ends of crevasses.

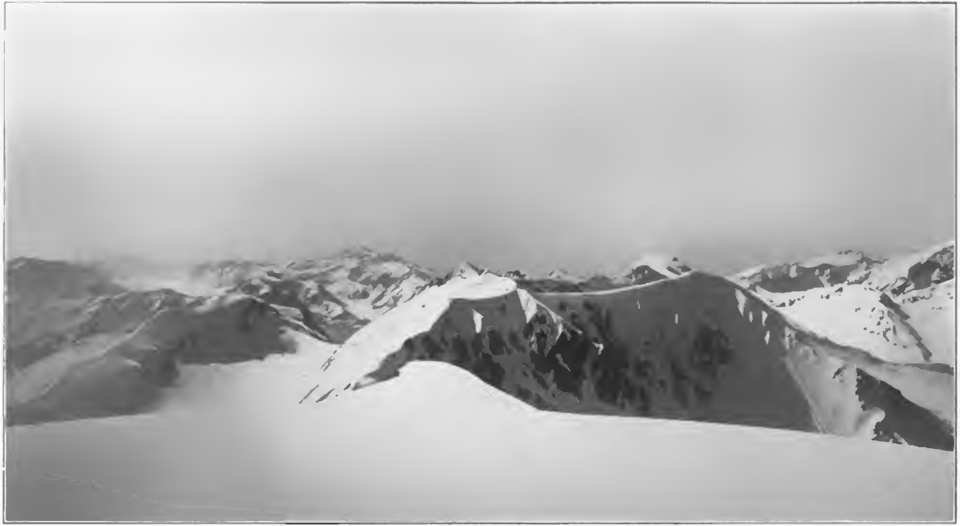
Along the margin, on both the east and west sides of the glacier, the condition of crevassing was widely different from that of the center. Here the ice was confusedly broken,

[1] Tarr, R. S., Professional Paper 64, U. S. Geol. Survey, 1909, p. 80.

A. Yakutat Bay from Mountain on West Side
Showing fringe of outwash gravels from Galiano and Atrevida Glaciers.

B. Terminus of Yakutat Glacier, 1906
Photograph by Fremont Morse, Canadian Boundary Survey.

PLATE II



VIEW LOOKING EAST FROM NORTH END OF PUGET PENINSULA

Showing subequality of the mountain peaks, but much diversity of elevation; heavy burden of snow and great snowfields; in the distance, on the left, the valley deeply filled with ice of the Hidden Glacier. Photograph (1896) by A. J. Brabazon, Canadian Boundary Commission. (From U. S. Geological Survey, Plate V, Professional Paper 64.)

VIEW LOOKING EAST UP THE TWO BRANCHES OF NUNATAK GLACIER FROM CREST OF THE NUNATAK

Photograph, July 5, 1906; it was possible to travel to the Alsek River valley across the broad ice divide. (From U. S. Geological Survey, Professional Paper 64.)

PLATE IV

VIEW LOOKING NORTH FROM PUGET PENINSULA

Russell Fiord on the right, bordered on the north by the straight slate coast. Showing rough subequality of mountain peaks and rise of the crests in the crystalline belt to the north; also vast extent of snowfields. East arm of Hubbard Glacier in valley to the right of the center. Photograph (1890) by A. J. Brabazon, Canadian Boundary Commission.

Plate V

View Looking East through Russell and Nunatak Fiords, with Nunatak Glacier in the Distance

The smoothed glaciated slate coast of the north side of the northwest arm of Russell Fiord on the left. Photograph (1896) by A. J. Brabazon, Canadian Boundary Commission.

A. Dissected, Ice-Sculptured Gravel Benches on North Side of Nunatak Fiord
Shows also gullied higher-level gravels on steeper mountain slope. Note absence of vegetation. Photograph taken July 8, 1905.

B. Morainic Surface Veneering Overridden Gravels, West Side of Russell Fiord
Vegetation just advancing over the surface from which the Nunatak-Hidden Glacier has recently receded, 1905.

MODEL OF THE MALASPINA GLACIER AND YAKUTAT BAY
Copyright, 1909, by the University of Wisconsin.

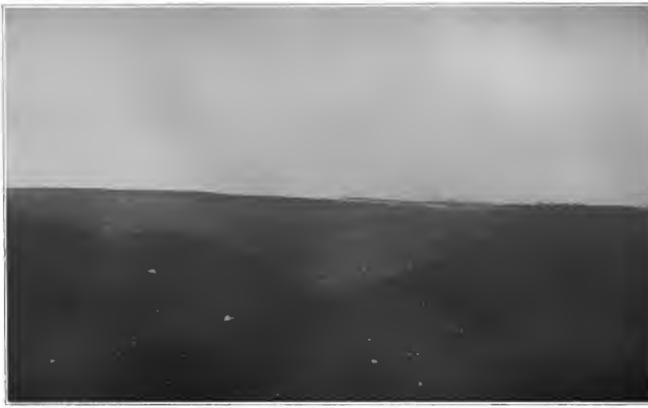
A. YAHTSE RIVER FROM ABOVE ICE TUNNEL
Photograph, 1891, by I. C. Russell.

B. YAHTSE RIVER ISSUING FROM A TUNNEL IN THE MALASPINA GLACIER
Photograph, 1891, by I. C. Russell.

A. Sitkagi Bluffs, on the South Margin of Malaspina Glacier
Photograph, 1891, by I. C. Russell.

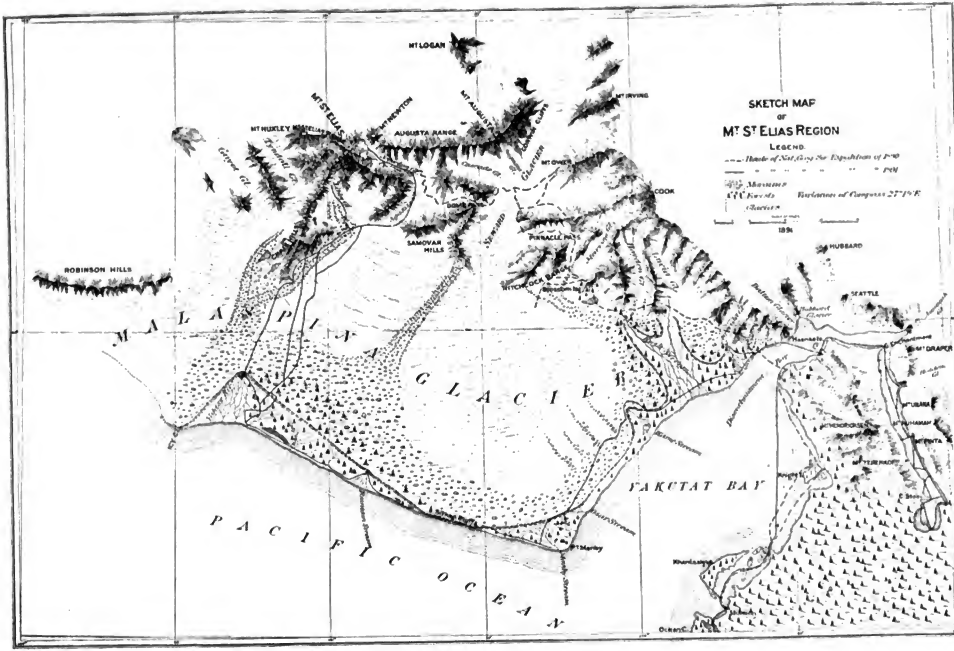
B. Central Portion of Malaspina Glacier
Photograph, 1890, by I. C. Russell.

PLATE X



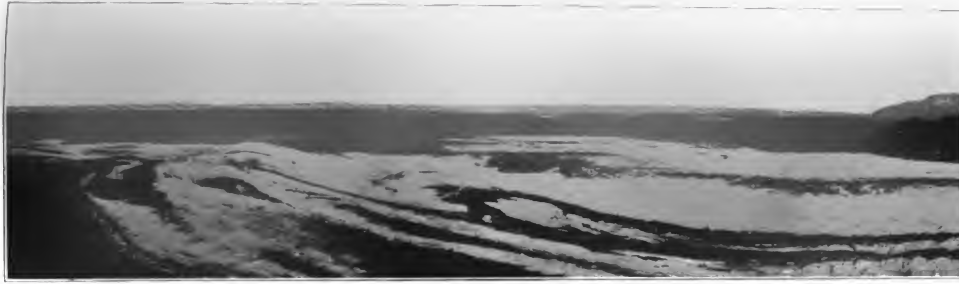
A. MORAINE-COVERED SURFACE OF MALASPINA GLACIER
Photograph, 1891, by I. C. Russell.

B. FOREST COVERING OF MALASPINA GLACIER
Photograph, 1891, by I. C. Russell.


SKETCH MAP
OF
MT ST ELIAS REGION
LEGEND
MT LOGAN
MT ST ELIAS
MT HUXLEY
MT NEWTON
AUGUSTA RANGE
MT AUGUSTA
MT IRVING
MT OWEN
COOK
SAMOVAR HILLS
SEWARD
GLACIER
PINNACLE PASS
HITCHCOCK RANGE
ROBINSON HILLS
HUBBARD
SEATTLE
MT DRAPER
MALASPINA GLACIER
YAKUTAT BAY
PACIFIC OCEAN
1891

A. Tyndall Glacier
A western tributary of Malaspina Glacier. Photograph from Chaix Hills by I. C. Russell, 1891.

B. Libbey and Agassiz Glaciers Joining the Malaspina Glacier
From Chaix Hills. Photograph by H. G. Bryant.



A. LOWER HAYDEN GLACIER

From elevation of 1585 feet on western side of Floral Hills. Moraine-covered Marvine lobe of Malaspina Glacier in background, August 22, 1905.

B. UPPER HAYDEN GLACIER

From elevation of 1585 feet on western side of Floral Hills. Looking up the glacier, Mount Cook on right. Large dark patch is the débris of a recent (1899?) avalanche. Photograph, August 22, 1905. The two pictures make a continuous panorama.

Plate XIV

A. BROKEN EASTERN MARGIN OF THE MARVINE LOBE OF MALASPINA GLACIER

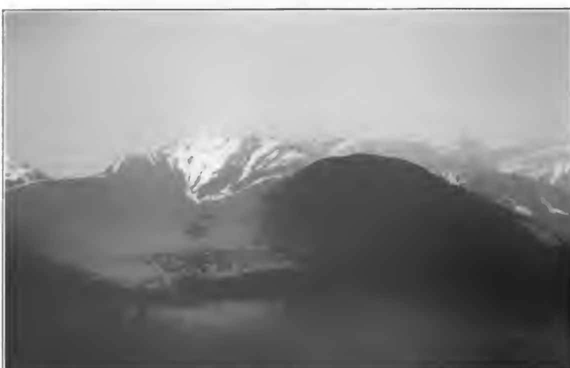
From alluvial fan of Kwik River. The broken ice cliff has protruded through the morainic soil and the forest. Blocks of ice fell from this cliff as the survey party passed, trees crashed down, and the morainic soil was constantly sliding down. Photograph taken August 11, 1906.

B. BLOCKS OF ICE PROTRUDING THROUGH MORAINIC SOIL AND FOREST COVER

Eastern margin of Malaspina Glacier. Cottonwood trees, in full leaf, overthrown by the recent thrust that broke the glacier margin. Photograph taken August 11, 1906.

A. MORAINE-COVERED FORESTED EASTERN MARGIN OF THE MARVINE LOBE OF MALASPINA GLACIER

The ice, recently thrust forward, has protruded through the soil; the trees are tilted at various angles and overturned; and the underlying ice, exposed to air and rain, is rapidly melting. Streams of water and of liquid mud descend the slope. In early July the foreground was occupied by a lake upon which the alluvial fan had encroached, the stream, in the meantime, being greatly increased in size. Photograph taken August 9, 1906.



B. LUCIA NUNATAK

With stagnant, moraine-covered ice riding up on its northern slope. Photograph, August 22, 1905, from Photo Station E (Map 2), at eastern end of Floral Pass.

the crevasses being short and extending in various directions, giving the glacier the appearance characteristic of an ice fall. Its pinnacled condition was only in small part due to greater ablation, for even here a large part of the surface was still protected by moraine. The glacier surface had the appearance of having been broken by dragging and shoving as the strain of the glacier flood swept down the deeper central portion and caused a swelling of the ice against the valley margins, the thinner marginal portion having too great rigidity to accommodate itself to the strains thus induced. One was reminded somewhat of the broken blocks of ice that are piled up along the margin of a frozen river that rises in flood. This condition extended all along the visible margin and was strikingly developed both on the stoss slope of the nunatak and around its eastern base.

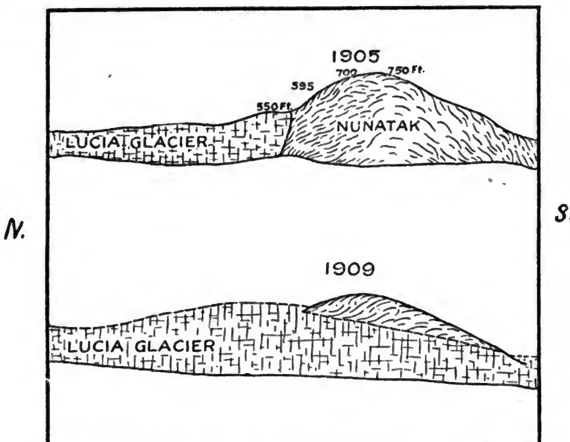


Fig. 4. Longitudinal Sections of Lucia Glacier and Nunatak.

Above Terrace Point the glacier was also badly crevassed. Those tributaries that enter from the west were not notably broken, but two that enter from the east, just above Terrace Point, were impassably crevassed where they emerged from their valleys, and the crevassing of these extended for only a half mile into the main glacier. It seems certain that some of the impulse of the advance of Lucia Glacier was supplied by these tributaries, but the greater part was apparently supplied from the main glacier, which was badly crevassed above these two tributaries, and, in fact, as far into the mountains as we could see. A noteworthy feature of the glacier above Terrace Point was the presence of a crescentic depression between the crevassed area of the two eastern tributaries and the crevassed main glacier. It is as if the glacier level were raised by the impulse from the main glacier, while the advance of the two tributaries had raised a crevassed

piedmont bulb where they emerged from their valleys; but between the two was an area that was less raised and broken, being between two different currents of recently accelerated ice motion.

Below Terrace Point the crevassing diminished in amount and finally died out completely before the alder-covered portion of the ablation moraine was reached. This outer portion, beyond the area of crevassing, was apparently in no notable respect different from its 1906 condition. Indeed three lakes showing in the forest in the 1906 photographs were still present on the glacier. There was no crescentic crevassing, such as was so prominently developed in Atrevida Glacier during its advance, and no noticeable thickening of the piedmont portion. This section was still a field of ablation moraine, with the outer fringe of forest and alder grading into a waste of barren, undulating, rocky moraine with no ice appearing when viewed from a distance. But in 1909 this desert of ablation moraine graded into an area of gashes in which ice was revealed in increasing amounts up the glacier.

While there was no noticeable thickening of the piedmont portion of the glacier, there was thickening along the margins at Terrace Point and Lucia Nunatak, and presumably also between these two points. The thickening and advance was best shown at Lucia Nunatak. Here the ice rode high up on the stoss or northern end. The undulating, moraine-covered stagnant ice that in 1906 rested high up on the northern slope of the nunatak was now replaced by an up-domed area of greatly crevassed ice which had overridden the barren area and advanced into the zone of vegetation that covers the nunatak top. At this point, therefore, the glacier was over 100 feet higher than in 1905 and 1906. A peculiar dome of broken ice also rose against the middle of the eastern side of the nunatak, but the cause for this could not be determined at the distance from which we viewed it. The stagnant tongue of Lucia Glacier which formerly stretched down the valley between the nunatak and Floral Hills, and which in 1905 and 1906 was not visible from the crest of Amphitheatre Knob, had been pushed forward and broken into impassable condition so that it was now plainly visible from that viewpoint, and the ice tongue extended a quarter of a mile beyond the southern end of the nunatak. Its total advance is estimated to have been between half and three-fourths of a mile. Whether it united with the main glacier below the nunatak could not be determined, but if it did not actually unite it certainly almost did so, and the hill was once more nearly if not quite a true nunatak.

At Terrace Point the ice had been pushed up laterally and, as already stated, had been greatly broken. During the earlier visits, in 1905 and 1906, this margin was a moderately-sloping, moraine-covered embankment with almost no ice showing, and one could ascend the slope at any point. Now the broken ice was bordered by a jagged precipice from which both moraine and ice blocks were constantly falling. The marginal valley between the glacier and Terrace Point had been greatly narrowed, the ice having pushed out laterally one or two hundred yards. At no point was the ice in contact with Terrace Point and a narrow marginal valley still existed. In the lower portion of this valley stood a narrow linear lake at least a half mile in length and two or three hundred yards wide, with the jagged ice precipice forming one margin and the southern end of Terrace Point the other. This lake was filled with icebergs, and as we looked down upon it we saw many fall into it. At the southern tip of Terrace Point in 1906 a small lake stood where the advancing Atrevida Glacier had pushed forward to dam the drainage from

the eastern margin of Terrace Point, forming a lake which did not exist in 1905. This lake still existed August 3, 1909, but south of it for over half a mile the moraine-covered depression between Lucia and Atrevida Glaciers was covered with icebergs. The expanded lake in which these icebergs floated was seen from the crest of Amphitheatre Knob on July 21, 1909, but before August 3 was drained through a subglacial tunnel.

Another notable change along the eastern margin of the Lucia Glacier was the development of a huge glacial stream in the depression between the outer portions of the Lucia and Atrevida lobes. In 1906 there was only a small stream here, but in 1909 one of the largest glacial streams of the region was flowing. This increase in volume was probably due in part to the recent breaking of the glacier, thus exposing the ice to more rapid melting; but this does not seem to be a sufficient cause for so great an increase. It is probable that a still more efficient cause is the interference with and destruction of the subglacial drainage which had developed during years of stagnation, and which in 1905 and 1906 found escape at the southern end of Lucia Nunatak on the opposite, or western, side of the glacier. The advance and breaking of the glacier, so notable near the nunatak, must have destroyed this system of subglacial drainage toward which most of the water from the glacier formerly flowed. Whether the new drainage all went to the east side, or whether a considerable portion still emerged on the western margin could not be determined; but that a large proportion escaped through the eastern stream is certain.

This remarkable change in condition of Lucia Glacier in the interval between August, 1906, and August, 1909, had wholly altered the appearance of the glacier, and yet we conclude that the thrust by which the change has been caused was either a weak one or else only just beginning to make its efforts felt, probably the latter. The reasons for this conclusion are several, all based upon the changes which a similar advance brought about in other glaciers. Compared with the Atrevida in 1906, for instance, the Lucia in 1909 was far less broken, and the piedmont area was not noticeably thickened, as the Atrevida was. As yet there had been no development of great crescentic crevasses in the piedmont bulb as was the case in the Atrevida, and the crevassing did not reach out into the alder zone as it did in the Atrevida. In other words, the stage of breaking of Lucia Glacier was in 1909 far less advanced than that of Atrevida Glacier in 1906. It was far less broken than the Marvine, Haenke and Variegated Glaciers also.

In view of the rapidity with which these waves of advance pass down a glacier, essentially the whole transformation in the case of those that advanced in 1906 having taken place within a period of nine or ten months, it is not improbable that the lesser development of the Lucia advance was not due to weakness of the thrust but to the fact that our observations were made too early and before the advance was finished. We can state with certainty that the entire breakage was recent, and mainly, if not entirely, in the season of 1909, and that it was still in progress in August. The proof of this is conclusive and of several kinds. Most noticeable of all is the condition of the crevasses and inter-crevasse or serac areas in the middle of the glacier. Although well below snow-line, and in a situation where ablation is rapid, the edges of the crevasses were sharp and angular as if freshly-broken, and the serac areas were still so completely covered with ablation moraine except on the crevasse faces that little ice was seen in this entire area. Even near the glacier margin, where the ice was so broken and pinnacled, a large part of the

ice was still veneered with moraine, a condition which could not possibly have been the case had ablation been long active on this maze of jagged pinnacles.

The marginal lakes, and the formation and drainage of one of them, point to the conclusion that the advance was in progress; but more important than this is the fact that during our visit the ice along the glacier margin was being broken by the thrust. As we stood on Terrace Point for a few hours we heard ice falls every few minutes, and that this had been in progress for some time was proved by the presence of many white ice patches in the broken, débris-covered marginal area, caused by the recent fall of ice, and even more noticeably by the great numbers of icebergs floating in the marginal lake and stranded on the bottom of the abandoned lake.

We believe that these facts demonstrate that the advance of Lucia Glacier began no earlier than the autumn of 1908 (and probably even more recently), and that it was in vigorous progress in August, 1909. We have no reason to doubt that in a few weeks a notable difference would have been observed in the appearance of Lucia Glacier. We looked forward with the utmost interest to re-examining this glacier in 1910, for it was to be expected that great changes would take place in the interval. If they had it would have given basis for a fuller statement of the nature of the changes in these advancing glaciers than is now possible. Hitherto the advance has been nearly or completely finished when observed, but in this case there is reason for thinking that the advance, as seen in 1909, had not reached the maximum. The ideal plan, had it been possible, would have been to stay right at Lucia Glacier and record the daily changes throughout the fall and winter.

Condition in 1910. When the junior author returned to Yakutat Bay in June, 1910, he regarded the observations of Lucia Glacier as the most interesting and important which it was his privilege to undertake. On going up Yakutat Bay an enormous amount of floating ice was encountered and, as it was then apparently impossible to cross to the west side of Yakutat Bay, the investigation of the Hubbard, Nunatak and other glaciers was undertaken first, in the hope that the ice jam would thin sufficiently in the course of a week so that the west side of the bay and Lucia Glacier might then be reached.

Upon the return to outer Yakutat Bay, however, the ice jam was undiminished but three attempts were made to penetrate through the icebergs to the west side of the bay, each time without success. A camp was maintained on the east side of the bay and the ice was watched constantly for favorable conditions but each time we attempted crossing, the launch was turned back by the ice jam, after some arduous and dangerous experiences. Finally, the date of the northward-bound steamer approaching, it was necessary to choose between not seeing Lucia Glacier in 1910 or losing two weeks or a month of the season in Prince William Sound, where the author had been directed to spend most of the summer. With deepest disappointment the Lucia Glacier observations were then given up.

Such observations as could be made with field glasses from high points on the east side of Yakutat Bay revealed, however, that Lucia Nunatak had not been completely overridden and that the stagnant moraine-covered outer bulb of Lucia Glacier had not been broken up. The ice rode little, if any, higher up on the north side of the nunatak than in 1909. The crevassed glacier surface was still dark with large amounts of ablation moraine upon the serac tables. The trees upon the visible outer part of the glacier bulb were undisturbed. The Kwik River and smaller streams north of it that receive dis-

charge from Lucia Glacier seemed not appreciably larger than in 1909, although as the observations were made fully two weeks earlier than the junior author had ever been in Yakutat Bay before, and as snow still lay on the ground at sea level, the latter point could not be stated with certainty.

These scanty and tantalizing observations gave the impression that the advance of Lucia Glacier had not continued very long after our leaving the region ten months before or else that the rate of advance had slowed down so that changes were taking place very slowly indeed.

Stagnation in 1911. By the summer of 1911 the advance of Lucia Glacier was all over, for it was easily crossed by a Boundary Survey party in that year. The return to stagnation after the advance of 1909 was evidently quite as rapid as in the other glaciers of Yakutat Bay which have had advances stimulated by earthquake avalanching. It was partly covered by ablation moraine when seen by the junior author from the bay in 1913.

Atrevida Glacier

General Description. Atrevida Glacier, which is much shorter than the Lucia, is an ice tongue of similar characteristics. It expands much more in its lower course, however, being only about a half mile wide in its deep mountain valley but flaring out rapidly to a mile and three quarters at the point where it passes between Amphitheatre Knob on the east and Terrace Point on the west, and coalesces with Lucia Glacier. The whole glacier, piedmont bulb and all, is probably less than ten miles in length. Several streams flow from it (a) to Kwik River, (b) directly to Yakutat Bay, and, (c) by Esker Stream, from the eastern side at the base of Amphitheatre Knob (Map 2).

Condition in 1890 and 1905. The glacier was named by Professor Russell in 1890,[1] who described it in about the same terms as the Lucia, besides showing its condition by several pictures. The 1890 conditions seem almost exactly like those in August, 1905, when we obtained a bird's eye view of it from the summit of Amphitheatre Knob, which rises above its eastern margin, and later in August when one expedition was made in various directions over the ablation moraine near the east side and another, by the junior author, from the east side to Terrace Point and back by different routes.

The exact length of the glacier is not known, but the valley portion is probably less than five miles, while the distance from Terrace Point to the terminus of the piedmont bulb is about five miles more. There are numerous tributaries, all small, so far as seen, and descending in steep courses as cascading glaciers. In the lower portion of the valley there are cascading glaciers which no longer unite with the main glacier. The valley walls, which attain elevations of 5000 to 6000 feet, are everywhere steep, showing clear evidence of profound valley deepening by glacial erosion, and between them lies the glacier with a width of two or three miles, broadening toward the valley mouth. So far as observed in 1905, with a single small exceptional area, Atrevida Glacier was in a semi-stagnant state in its valley portion and quite stagnant on its outer portion beyond the mountain front. It was possible to travel easily on any part of the glacier and we walked freely over its surface. For a mile or a mile and a half within its mountain valley the glacier was covered with an almost continuous sheet of ablation moraine and the surface was an undulating waste of rock fragments of all sizes, from huge angular bowlders to

[1] Nat. Geog. Mag., Vol. III, 1891, pp. 92–5, and Plates 10 and 11.

clay. The only difficulty in travelling and packing camp outfit over this surface was the unstable position of the rock fragments resting on the glacier, and the necessity of constantly climbing ridges and knobs and descending into pits and valleys, in whatever direction one travelled. There seemed to be no system to the hillocks and depressions, their origin being evidently differential ablation of the under ice which the moraine protected in varying degrees according to the local thickness of the moraine cover.

In the upper portion of this morainic desert the amount of visible ice increased, the moraine cover became notably thinner, and the surface was less irregular. Some crevassing also appeared, but nowhere in sufficient amount to prevent travelling over the glacier except in one small section where a dome of crevassed ice appeared in the midst of the moraine desert. It was so abnormal a feature that we photographed it, thinking that it represented the updoming of ice in passing over some elevation in the valley bottom. We predicted the existence here of an unborn nunatak, but the crevassed dome is now interpreted as the first sign of the coming transformation of the glacier which was observed in full progress in 1906. Still farther up the glacier, clear ice predominated and finally the moraine entirely disappeared, but we did not visit this part of the glacier and, therefore, cannot describe it in detail.

Nowhere in the upper part of the valley portion of Atrevida Glacier did vegetation grow in the ablation moraine, for it was shifting rapidly. But farther down, near the end of the valley portion, scattered alder bushes and clusters of alder grew, extending farther up the valley on the eastern margin than in the center, because the lateral moraine deposits were thicker and, therefore, under more stable conditions. Just outside the mountain valley portion of the glacier the ablation moraine was covered with alder, and farther out, with cottonwood and spruce. These bushes and trees were mature, indicating a long condition of stagnation in this part of the glacier. The stagnant outer part of the Atrevida was a true piedmont bulb, extending four or five miles beyond the mountain front and expanding fan-shaped until it attained a width of about twice that of the valley portion. On the western side it coalesced with Lucia Glacier. As already stated in the description of that glacier, the two glaciers were separated in the lower portion by a V-shaped valley, then by an interlobate section on which a narrow strip of alder grew, and above this the two glaciers again separated a short distance below Terrace Point, which, with the mountain back of it form the dividing walls between the valley portions of the two glaciers.

Both at Terrace Point and along the eastern side of the valley portion of Atrevida Glacier the margin had a thickly moraine-covered, moderately-sloping embankment which we could ascend at any point. It formed one wall of a marginal valley, carrying some drainage. On the west side there was little drainage from either the Lucia or Atrevida, and only a small stream escaped in the V-shaped valley between the interlobate portion of the coalesced Lucia and Atrevida; and there were only small streams emerging from the outer portion of the piedmont bulb which rose above the Kwik valley. A small stream emerged from the forest-covered part of the piedmont bulb on the eastern side and flowed into Yakutat Bay, but it was easily forded. The main drainage of the glacier emerged from the east side at the base of Amphitheatre Knob. At the point of emergence of this stream, which Russell called Esker Stream, there was a low cave with vertical or overhanging walls above which the ice edge had a slope of about 35°, down which a few stones were sliding. The ice was nowhere heavily débris-laden, though dirtiest

near the base. There were a few crevasses, dipping like joint planes at an angle of 5° or 10° from the horizontal and suggesting differential ice movement. At the top of the 200 foot ice slope there was a dense alder thicket with bushes estimated to be fifteen or twenty years old. We are inclined to interpret the slight crevassing of the ice cliff as the beginning of the advance and profound breaking which were observed nine months later.

Professor Russell crossed Atrevida Glacier in 1890, starting at the point of emergence of Esker Stream, with his entire outfit. From his description we are led to infer that the Atrevida was in essentially the same condition in 1905 as in 1890. One of the photographs illustrating his report[1] shows the glacier surface covered with ablation moraine; another shows the ice cave from which Esker stream emerges and we could notice scarcely any difference in its form in August, 1905. Russell's description of the conditions in 1890 was as follows[2] "The waters, brown and turbid with sediment, welled out of a cavern at the foot of an ice precipice 200 feet high, and formed a roaring stream too deep and too swift for fording. . . . The dark-colored ice, mixed with stones and dirt, might easily be mistaken for stratified rock, but the dirt discoloring the ice is almost entirely superficial. The crest of the cliff is formed of débris, and is the edge of the sheet of stones and dirt covering the general surface of the glacier. Owing to the constant melting, stones and boulders are continually loosened to rattle down the steep slope and plunge into the water beneath."

There is every reason to believe that the glacier had changed very little for many years before Russell's visit. Its piedmont portion must have been in a stagnant state for at least a half century before 1905, otherwise the ablation moraine and its forest could not have developed. That it had not been notably more expanded for an even greater length of time was proved by the presence of a mature forest growing up to the very margin of the moraine-covered ice. No notable recent recession had taken place, for there was only a very narrow area along this margin in which the forest did not grow. From the evidence that we possess we feel warranted in inferring a long period of stagnation, with no notable advance or recession for several decades, and probably for more than half a century. Up to the autumn of 1905 the Atrevida was a fine example of a valley glacier with a piedmont terminus completely covered with ablation moraine and in the outer portion so stagnant that it bore a dense growth of vegetation. No one would have expected that such a glacier would suddenly spring into activity, and such a thought never occurred to us even when, in June 1906, the senior author approached the Atrevida with the purpose of traversing a route westward to the Malaspina along the line followed by Russell in 1890 and by the junior author in August, 1905.

The Advance of 1906. As the Geological Survey party approached the margin of Atrevida Glacier early in June, 1906, never dreaming of change, the first surprise came when at a distance of several miles a jagged ice cliff was seen where nine months earlier there was a low, moraine-covered ice margin, bearing forest; and later investigation showed that the entire glacier, with the exception of the outermost portion, was utterly transformed and quite impassable. A bird's-eye view of the entire glacier was obtained from the crest of Amphitheatre Knob (Pls. XXVI, XXVII), which rises directly above the eastern margin of the Atrevida, and early in August a visit was made, via Floral

[1] Russell, I. C., An Expedition to Mount St. Elias, Alaska, Nat. Geog. Mag., Vol. III, 1891, Pl. X.
[2] Nat. Geog. Mag., Vol. III, 1891, pp. 94–5.

Pass, to the western margin at Terrace Point. Such a sudden and absolute change in a glacier in so short an interval of time had never before been observed and described, so far as we know.

From side to side, and from as far up the mountain valley as we could see, down into the alder-covered portion of the piedmont bulb, was such a sea of crevasses (Pl. XX) that there was no direction in which it seemed possible to cross the glacier; and even to ascend to its surface proved to be a matter of great difficulty. Where formerly there was a waste of ablation moraine, with no ice appearing in a distant view, the glacier was so profoundly broken that clear ice appeared on every hand. Much of the ablation moraine had disappeared into the crevasses and during the summer still more was thus removed, so that the glacier surface was transformed not only by the development of innumerable crevasses but also by the loss of much of its morainic veneer. Moraine still clung to the tops of the broader serac ridges, but more than half the surface was clear ice; and instead of the undulating morainic surface there were ice splinters, pinnacles, arêtes, and profound crevasses extending in all directions.

In the valley portion of the glacier there was no appearance of system in the crevassing, but beyond the mountain front, where the glacier spreads out, the crevassing assumed a crescentic form. A series of concentric gashes, with radius increasing toward the outer portion of the crevassed area, showed their crescentic form very clearly where the pure ice was opened in the dark ablation moraine, and still more clearly in the dark-green alder zone. It was almost weird to see these ice-walled gashes rent in the soil in which dense thickets were growing, but it showed vividly on what an unstable foundation the vegetation was growing. Toward the outer margin of the broken area the crevasses gradually died out, the outermost ones being short, narrow rents. Ablation was proceeding with rapidity, the soil beneath the plants was sliding down the slopes and into the crevasses, and they were being undermined and overturned. It was evident that the alder thickets on that part of the glacier which was broken were doomed to destruction; but whether the ice would be still further broken here, or whether the area of crevassing would be extended farther into the forest-covered part of the glacier could not be foretold.

Not less remarkable were the accompanying changes along the glacier margin, on both the east and the west sides. At Terrace Point, on the western side, the ice was crowding up on the land, overriding the gravels and pushing them up in low ridges. Below Terrace Point the ice had pushed out some distance, in the low interlobate area between the Lucia and Atrevida Glaciers, overriding a camp site occupied in 1905, and forming a new lake by the extension of an ice dam across the marginal drainage of the two glaciers. The ice was also noticeably thickened; for in 1905 the Atrevida had about the same elevation as the Lucia where the two coalesced, the boundary between the two being marked by a strip of alder; but in 1906 this alder was destroyed by lateral thrust and the area of crevassed Atrevida ice was much higher than the ablation moraine of the adjoining stagnant Lucia. The thickening amounted to fully 200 feet near the boundary between the two glaciers. At the same time the crevassed Atrevida bulb extended across the interlobate section into the area of the Lucia, and, being both higher and much crevassed, it was then easy to trace the boundary between the two glaciers.

A noteworthy feature of the glacier just below Terrace Point, and thence eastward three-quarters of the distance to Amphitheatre Knob, was a roughly-crescentic area of white ice bordered on both the up-stream and down-stream sides and on both margins by

débris-covered ice. The surface of the white area was crevassed as badly as that of the surrounding débris-covered glacier. The appearance of this area of white ice (pp. 78–9) in the midst of moraine-covered ice, and where débris had previously covered the glacier, was a puzzling feature. The significance of this feature is discussed in another chapter (p. 187), after other similar areas have been described.

The view of the broken glacier from Terrace Point (Pl. XIX, A) showed clearly that in 1906 Atrevida Glacier was impassable from as far up the mountain valley as we could see down to the outer border of the crevassed area in the alder zone. This view, over a month later than the one from Amphitheatre Knob, showed considerable change in detail through ablation, but otherwise no notable difference in condition. The western margin of the glacier near Terrace Point consisted of a high, crevassed, inaccessible ice precipice replacing the moderately-sloping, moraine-covered embankment, up and down which we easily traveled in 1905. By cutting ice steps all the way this margin could be ascended; but further progress was barred by a complex network of profound crevasses, with intervening ridges and pinnacles, that could not be crossed with a camping outfit. It was a wonderful contrast to the easily-traversed surface of 1890 and 1905.

Along the eastern margin titanic changes, quite like those along the margin of the advancing Marvine Glacier were in progress before our very eyes. At Esker Stream the ice front had advanced an unknown amount, destroying the vegetation on the slopes of Amphitheatre Knob. The moderately sloping cliff of 1890 and 1905 was changed to a jagged precipice of broken ice blocks, down which débris was incessantly falling. The cave from which Esker Stream issued had disappeared through ice faulting (Pl. XXI, A), and the stream had increased in volume. The faulted, splintered ice cliff extended westward with overhanging cornices due to horizontal thrust faulting (Pl. XVII, B). In the one place where the cliff could be ascended, further advance was halted by deep crevasses before the level of the glacier surface was reached. The morainic soil, with the bushes growing upon it had disappeared into crevasses or had slid down the front of the glacier. Moraine, ice, and wood were constantly sliding down from the glacier front, the avalanches sometimes including score of tons, and it was always dangerous to the observer. By this material the coniferous forest growing in front of Atrevida Glacier was being tilted, buried, and destroyed, including alders, underbrush, and trees over 50 years old (Pl. XXII, B). The breaking also resulted in exposing to melting the buried ice that had long been protected by moraine and forest, so that augmented streams were rapidly cutting into the soil, gravels and rock. All this change had taken place between August 23, 1905 and June 26, 1906.

Condition in 1909. Knowing of no previous observations upon the behavior of glaciers undergoing abrupt change from long continued stagnation to great activity, and from unbroken surface to impassably crevassed condition, we could not, in 1906, predict with any certainty what was in store for the Atrevida Glacier. In view of the suddenness of the transformation, and the rapidity with which its full effects spread over the glacier, we were prepared to believe that the cycle of activity would be a brief one if the cause of earthquake impulse were the correct explanation. We did not, however, suspect what the evidence of the 1909 observations clearly demonstrate, that the advance had passed its height and had almost ceased when the Atrevida Glacier was being studied in the summer of 1906. The probable future of Atrevida Glacier predicted in 1906 by the senior author was that the alder and forest growth within the crevassed area would

be destroyed, with perhaps a larger area if the forward thrust reached further. It seemed probable that the latter would be the case, for the breaking observed in 1906 suggested that the maximum had not been reached. As a matter of fact the Atrevida Glacier had practically ceased its forward motion in August, 1906, and the termination of its advance, and its return to a condition of stagnation, have been almost as spectacular as was its transformation into a state of activity between August, 1905 and June, 1906.

In 1909 we reoccupied the sites from which the observations and photographs of 1905 and 1906 were made, and were, therefore, able to make exact comparisons of the condition in the several years (Pls. XX, and XXI, A). No new observations of Atrevida Glacier were made during the two weeks spent in Yakutat Bay in 1910. In 1909 it was clear that the glacier had advanced but little since last seen in August, 1906, that the wounds from the spasmodic advance had been to a great extent healed by ablation, that the glacier had again relapsed into a state of stagnation, and that it was now once more possible to travel over its surface. In 1909 the Atrevida Glacier was far more like what it was in 1905 than in 1906, but its surface was much rougher than in 1905. The greater part of the glacier was covered with a sheet of ablation moraine (Pl. XIX, B) and this extended much farther up the valley than it did in 1905. This is a remarkable fact, and no other explanation occurs to us than that the extension of moraine cover up the glacier valley is due to the downshaking of rock fragments from the steep valley walls during the 1899 earthquakes, thus furnishing an unusually great supply of débris that was above the snow line in 1905 but had been brought down into the zone of ablation by the rapid advance of 1906. A few small areas of clear ice appeared in the waste of moraine that covered the glacier above Terrace Point, most of them in the middle of the glacier, where one would naturally expect less débris from avalanches. This upper glacier surface was viewed and photographed at the same site on Terrace Point, in both 1906 and 1909 (Pl. XIX) and the photographs show clearly a decided lowering of the ice surface in the interval of three years. It is impossible to state the exact amount, but it cannot be less than 50 feet and may be even more than 100 feet. This lowering of the ice may not all be the result of ablation, for there may have been a flattening through the forward creep of the lower ice of the glacier after the spasmodic advance ceased.

Between Amphitheatre Knob and Terrace Point, just where the valley glacier portion may be said to end, more detailed studies were made than elsewhere, because it was here that Russell crossed in 1890, and the junior author in 1905, while it was carefully examined from both the east and west sides in 1906. There was, therefore, a better basis for comparative study here than elsewhere. In 1909, therefore, this portion was examined and photographed from Amphitheatre Knob and from Terrace Point, as in 1906, and it was crossed, as in 1905. With the exception of a narrow medial area, in which small patches of clear ice appeared, the entire surface was covered with a sheet of ablation moraine as it was in 1905; but the surface was much rougher than then, the morainic hillocks being higher and the intervening pits and valleys deeper. No sign of the crevassed dome observed in 1905 was visible in 1909. As compared with 1906 the surface was of course far smoother, and when viewed from either side no crevasses were visible, and, excepting in the area mentioned, almost no ice could be seen. This contrasts strikingly with 1906 when the glacier was broken by a maze of crevasses, and more than half the surface was clear ice; and the contrast is brought out most strikingly by the photographs taken from the same site on Terrace Point in the two years. In these

photographs, also, a distinct lowering of the ice surface is noticeable, in this case surely of more than 100 feet.

It was found possible to walk freely over the ablation moraine though, owing to its greater roughness, with much more difficulty than in 1905. Alder was no longer growing on this part of the glacier but there was a thick cover of ablation moraine, though with many bare, débris-stained slopes and others with only a thin veneer of moraine. Melting was in rapid progress and small streams were coursing down all the slopes, but there were no large moulins, for the ice drainage had evidently not yet progressed far enough to develop these as they were developed in 1905. Crevasses were seen in many places and care was necessary lest one should step into those hidden from view by the moraine. Several times such a crevasse was unexpectedly discovered when stepping on what seemed to be solid, moraine-covered ice. We soon learned that the ridges were the places to be avoided and that the depressions offered the safest routes, for few and only shallow crevasses were found in the valleys, while the ridge crests were almost uniformly the sites of crevasses. The explanation of this fact is not difficult. After the glacier was broken by the advance, the moraine slid into the newly-formed crevasses, and, when ablation proceeded to lower the glacier surface, these filled, or partially filled, crevasses, with their greater depth of moraine, were so protected that they were preserved and stood above the level of the inter-crevasse areas. Then the rock fragments on and in these elevated crevasse areas slid down into the valleys, and doubtless the next stage will be the rising of the filled valleys and the lowering of the emptied crevasse areas.

In addition to the ridges due to crevasses the ablation moraine is diversified by bands and patches of rock predominantly of a single kind, such, for example, as black shale or conglomeratic rock, and in these areas are found unusual numbers of large angular blocks. They did not seem to be arranged in any order that could be correlated with flowage lines and may possibly be interpreted as rock falls and avalanches transported down the valley and spread out irregularly by the movement of the ice and by distribution through sliding, as ablation lowers the ice and undermines ridges and fills depressions.

It would be interesting to know the rate at which ablation has been proceeding on this ice surface. That the rate must have been rapid is indicated by the evidence furnished by the Atrevida and other glaciers. This evidence is of several kinds. A comparison of photographs from the same point show a general lowering of the surface sufficient to be apparent in photographs. It is possible, as already stated, that some of this lowering may be due to flowage after the first vigorous advance, but much of it, is certainly due to ablation, for the old crevassing and the noticeable proportion of clear ice have both been in large part destroyed, and only ablation could account for this. Moreover, the transformation of crevasses to ridges 50 to 100 feet high can be explained only as a result of ablation.

The only specific knowledge that we have regarding the rate of ablation in this climate are the measurements made on Hayden Glacier in 1906 when in a period of twelve days, in late July and early August, it was found that the surface of the glacier was lowered at the rate of 4 inches a day. This, however, was on a smooth ice surface, and at a much greater elevation, where the temperature on clear days descended below freezing point even before the sun set. The surface of Childs Glacier was lowered 7 inches a day in

July, 1909. The surface of the Baird Glacier of southeastern Alaska was lowered an inch a day by ablation in 1893. Muir Glacier surface melted 2 inches a day in 1890 and 1892. On the broken Atrevida, much nearer the sea, and at a lower elevation, the rate of ablation must have been greater than that on the Hayden. But when the ice surface began again to be sheeted with moraine the rate of ablation must have notably decreased. With these variable factors, and the absence of exact measurements, we can do no more than make the general statement that ablation, at first rapid, then at a decreasing rate, had so greatly lowered the surface of the Atrevida Glacier between August, 1906 and July, 1909 as to partially heal the crevasses that resulted from the spasmodic advance of 1906 so that it was possible to once more travel over its surface.

Specific measurements of the thickening of Atrevida Glacier through the 1906 advance, and the subsequent thinning through ablation before 1909 are based upon a pair of cross-

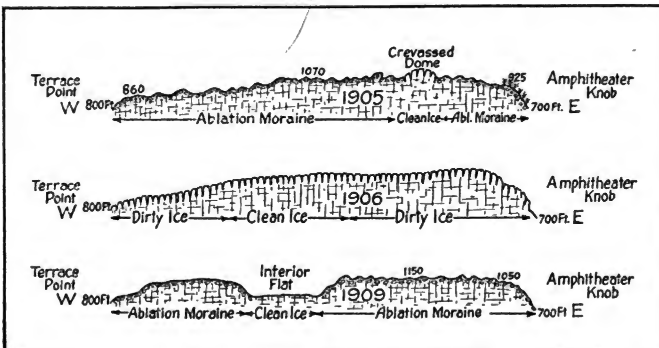


FIG. 5. CROSS-SECTIONS OF ATREVIDA GLACIER.

sections of Atrevida Glacier which the junior author made, one in 1905 before the advance had commenced, the other in 1909. The 1905 and 1909 elevations are barometric. These are shown in the three profiles of Fig. 5. The slopes and elevations of the 1906 cross-section are hypothetical. Each cross-section is from Esker Stream at the base of Amphitheatre Knob on the east side of Atrevida Glacier to Terrace Point on the west side. It will be seen that the 1909 cliff at Esker Stream is 125 feet higher than that of 1905. Assuming ablation here of 50 to 100 feet we should have had an ice cliff in 1906 of 400 to 450 feet. The thickening would therefore be 175 to 225 feet at least. In the middle of the glacier the thickening, similarly computed, must have been at least 130 to 180 feet. These comparisons, though in part based upon rough estimates, clearly indicate great thickening, and may safely be considered minimum measurements, since the amount of ablation is unknown, our calculation being based upon average depth of crevasses, to or near the bottoms of which ablation has now reduced the glacier surface.

The margin of Atrevida Glacier was examined both at and near Esker Stream and at

Terrace Point. In the latter place the glacier margin had lost its precipitous, broken ice cliff which was replaced by a moderately-sloping embankment of moraine-covered ice, steeper than in 1905, but ascended easily at practically all points. The margin stood almost exactly where it did in 1906 having melted back only a little, leaving a narrow marginal valley with no vegetation either in its bottom or for a short distance up the side of the terrace.

Along the eastern margin, near Esker Stream, the jagged cliff was also gone and one could again ascend the glacier (Pl. XXIII,) but up a much steeper slope, with thinner moraine cover than in 1905. The broken ice cliff above Esker Stream (Pl. XXI, B) was healed and the stream once more emerged from a well-defined ice cave (Pl. XXII, A) though a much smaller one than that which existed previous to the advance. The stream volume was greater than in 1905 but less than in 1906. It was clear to us, though the evidence as to the exact amount is not conclusive, that this part of the glacier continued to advance for a short distance after it was last seen in 1906. This is supported by photographs in the two years showing the trees at the ice margin just west of the stream, where the ice had advanced many yards into the forest.

Since the advance ceased there had been a slight recession, leaving a narrow barren zone like that at Terrace Point. At the base of Amphitheatre Knob there was a confused mass of broken and battered alder which was bombarded by falling bowlders and ice blocks during the advance; and in the débris that accumulated at the base of the ice cliff during this period, were incorporated the remnants of the alder thicket that grew on the glacier margin and in the marginal valley into which the ice cliff had been pushed. On the west side of Esker Stream, overturned and inclined spruce trees were seen in the morainic débris at the base of the glacier margin as they were in 1906; but instead of a jagged ice cliff rising above them (Pl. XXII, B), and constantly discharging bowlders, ice blocks, and torrents of muddy water and mud flows into the forest, there was a moraine-covered embankment down which small rock slides fell occasionally (Pl. XXIV, A). The destruction of this forest, so actively in progress during the summer of 1906, had ceased; but the effect of the advance on the fringing forest and on the alder thicket growing on the ice was strikingly apparent. There were hundreds of dead and dying spruce trees fringing the ice and partly buried in the morainic débris, and there were thousands of dead alder bushes, often in windrows, on the ice surface.

In the piedmont ice bulb that spreads out beyond the mountain base was seen evidence of the same changes as have already been described for other parts of the area. It is clearly evident that the advance had about ceased when observed in 1906 and that the outer border of the glacier was unaffected by the advance. It, therefore, still retained its forest of alder, cottonwood, spruce, and hemlock. In the broken area the crescentic rings of breakage were still plainly visible but ice no longer appeared in them and no crevasses were seen. Ablation had so healed the scars that, viewed from a distance, the entire surface was a waste of ablation moraine; but there was a system of concentric crescents of ridges and valleys developed by ablation in the formerly broken area. The extent to which ablation had proceeded here since 1906 was vividly brought out by the destruction of the alder thickets. In 1906 the alder still grew on the inter-crevasse areas, and the crevasses formed crescentic white gashes in a field of green. Now the alder was all dead within the broken area and it lay in brown windrows piled up on the moraine-covered ice.

Thus, proceeding down the glacier surface, one sees, first, barren ablation moraine with no vegetation, then ablation moraine with crescentic ridges and valleys littered with dead alder, and finally ablation moraine covered with dense alder thickets undisturbed by the advance. No evidence as to the extent of lowering of this part of the glacier was seen, but the photographs show clearly that the western portion of the piedmont bulb, as observed from Terrace Point, had been greatly lowered, and that a series of crescentic ridges and valleys had developed in the moraine-covered outer portion. The amount of lowering here is comparable to that observed elsewhere. The crescentic breaking, and the ridges and valleys which have succeeded it, present the appearance of viscous flowage of an underlying ice stream, carrying with it and breaking up a rigid upper crust of ice.

One of the most notable features on the Atrevida Glacier, as seen in 1909, was a large, roughly-crescentic area of clean white ice, with only a few short ridges of débris upon it, just outside the mountain front (Pl. XXV, B). Upstream from it lies the valley portion of the glacier which is almost completely covered with débris; and downstream from it lies the area of crescentic crevassing of the piedmont bulb, already described. This clear ice area extended westward nearly to Terrace Point, broadening in that direction and occupying a part of the bulb which protrudes toward Lucia Glacier—the part of the Atrevida in which there was the greatest thickening and forward movement. In 1905 the site of this area of clear ice was occupied by moraine-covered ice, but in 1906, (Pls. XXVI, XXVII) as already mentioned, it appeared in the broken glacier surface as an area of somewhat-domed, clear ice (Pl. XXV, A) surrounded by broken, partially-débris-covered ice. On its inner, or up-stream, side the glacier surface rose abruptly in 1909, as a high wall covered with moraine, and above this the glacier was almost completely covered with ablation moraine. The clear area was fully a mile and a half long (from east to west) and half or three quarters of a mile wide in the middle, which is the widest part, narrowing toward both the east and the west. It had spread somewhat in the interval between our photograph of 1906 and the period of observation in 1909. In 1909 moraine-covered ice rose above the clear area on the down-stream side and on both ends as well as on the up-stream side, but the change was less abrupt and pronounced in these directions than up-stream. The clear ice area formed a distinct depression in the glacier surface. Down-stream the clear ice area gradually died out, bands of débris appearing in it. The fact that this area was lower than the surrounding débris-covered ice is probably the result of more extensive ablation in the clear part of the glacier; and that there had been much ablation here is further proved by the fact that most of the crevassing, by which the area was severely broken in 1906, had disappeared. Many shallow crevasses, broadened by melting, still remained, but they were only the remnants of the former extensive gashes,—the crevasse bottoms. The surface was quite rough, but travel over it was not difficult, only short detours being necessary to avoid the crevasses (Pl. XXIV, B). Although not directly on the route it was chosen as the easiest path from Amphitheatre Knob to Terrace Point.

In seeking the explanation for such a peculiar area of clear white ice in the midst of a waste of ablation moraine we are aided by four noteworthy facts: (1) it has appeared as a result of the 1906 advance; (2) morainic débris that covered the ice on this site in 1905 had disappeared, for practically none remained even in the crevasse bottoms; (3) the clear ice area was distinctly domed in 1906; (4) some, at least, of the layers of ice in the

white area are highly inclined and in some cases vertical. All these facts harmonize with the hypothesis that at this point there was an upwelling in the glacier, bringing clear ice to the surface and pushing aside the débris-covered ice that previously stood here. We arenot able to postulate any other hypothesis which will explain these notable facts. The process which we conceive is, first, the fairly-free flowage of the ice down the mountain valley. Beyond the mountains further flowage was resisted by the stagnant, expanded bulb, which, being thin, was too rigid to easily yield to the pressure upon it. It yielded somewhat, however, and a series of crescentic gashes were opened in the piedmont area, but the resistance was so great that relief for the pressure from above was found in three other ways: (1) by an advance westward toward the most open area, (2) by a pronounced thickening and doming of this part of the glacier, and (3) by the uprising of ice from below.

The future of this area of clear ice seems to promise the development of an interior flat—a nearly moraine-free depression in the midst of ablation moraine such as is observed in the Variegated, Galiano, and other glaciers. The cause for these interior flats has hitherto been a mystery to us, but this hypothesis offers an explanation. Further consideration of this subject will be deferred to a later chapter after the characteristics of other, older, and better-developed interior flats have been studied. If this hypothesis is correct the phenomenon of interior flats must have an important bearing on the question of the mode of motion of glaciers. Many facts observed in this region of advancing glaciers point to viscous flowage as the mode of motion, but the discussion of this subject also will be deferred until other facts of observation have been stated.

CHAPTER V

THE GALIANO AND BLACK GLACIERS

Galiano Glacier

General Description. Although small, this is one of the most interesting glaciers of the Yakutat Bay region. It has been known since 1890 when Russell named and described it, and, although it has undergone some remarkable changes in the interval, its general features are now much as in 1890 (Map 2, in pocket).

Like many other neighboring glaciers, the Galiano consists of two quite different parts, a valley glacier portion and an expanded piedmont bulb beyond the mountain front; but to one unfamiliar with the phenomena of ablation moraines neither the bulb nor the valley portion looks like a glacier, for in all but a few places the ice is obscured from view by thick moraine, bearing alder on the outer portion. The valley which the Galiano Glacier occupies is broad, deep, and cirque-like with steeply-rising walls, and is enclosed in mountains which rise to elevations of 5000 to 6000 feet and bear a heavy cover of snow. From these snowfields several cascading glaciers descend (Pl. XXVIII), especially at the very head of the valley, supplying the ice for the glacier and a large part of the morainic débris which coats its surface; but some of this is also supplied by avalanches from the bare rock slopes which have been steepened by glacial erosion. The sound of falling stones is heard every few minutes when one is in this valley.

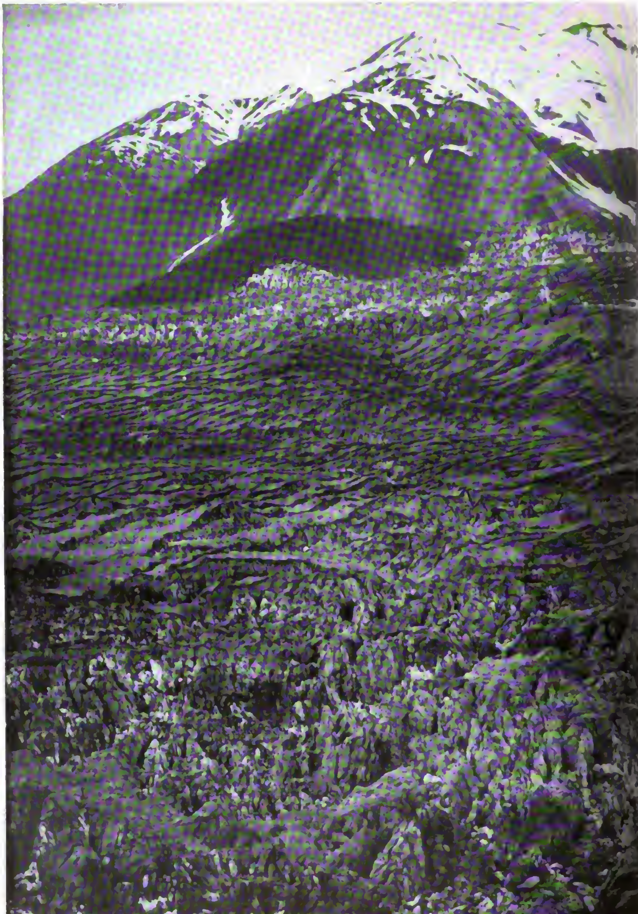
The cascading tributaries are a mile to a mile and a half long. The valley portion of the glacier is about three miles and the piedmont bulb is three quarters of a mile long. and a mile wide. The breadth of the glacier at the entrance to the mountain valley is three quarters of a mile while in the mountain valley the width is only a little over a quarter mile. With the steeply-rising snow-capped sides and head and the broad lower part, one is greatly deceived in the length of the valley when seen from a distance. It does not appear to be more than a mile or two long. The glacier has a slope from the mouth of the valley to its head of between 10 and 15 degrees. At no time when observed has this glacier been much crevassed, though in 1905 crevassing was observed near the valley head; and where the tributaries entered there were also areas of decided crevassing. Below its crevassed uppermost portion the Galiano Glacier has the appearance of stagnation, an appearance which finds ready explanation in the small size of the contributing glaciers. Indeed, it seems quite remarkable that such a limited ice supply could maintain a glacier of such length and breadth with its lower end nearly at sea level; and it is very doubtful if they could do so were it not for the fact that the wasting of a large part of the glacier is greatly retarded by the moraine cover which it bears.

Just within the mountain valley, on either margin of the glacier, there is a marginal valley with moraine-covered ice for one wall and the mountain rock for the other. In each of these valleys is a marginal stream, that on the west side being much the larger

Plate XVII

A. Looking up Lucia Glacier from Crest of Lucia Nunatak
Moraine-covered ice in foreground. Mount Cook in background. Photograph from Station D (Map 2), August 22, 1905.

B. Eastern Margin of Atrevida Glacier
Advancing into alder thicket at western base of Amphitheatre Knob. Photograph, 1906.



Lucia Glacier in 1909 from Terrace Point

It was crevassed and moving rapidly. In 1890, 1905, and 1906 it was stagnant and without crevasses. The nunatak in the background was being overridden. Photograph, July, 1909, from Station B (Map 2).

A. THE CREVASSED SURFACE OF ATREVIDA GLACIER
From Station C (Map 2), Terrace Point, August, 1906

B. ATREVIDA GLACIER
From same station as upper view, July, 1909. Again mantled with ablation moraine (as in 1905) and crevassing largely healed; but surface both higher and rougher than in 1905.

Plate XX

Long-Focus View of Crevassed Surface of Atrevida Glacier

[illegible] Station B (Map 2) Terrace Point. Photograph taken August 2, 1906. See also Plate XXI, A.

A. Lower Atrevida Glacier (August, 1909)

From same site as Plate XX. In the interval the crevassing has been healed and the mantle of ablation moraine has again covered the surface, so that, though both higher and rougher, the condition approaches that of 1905.

B. Esker Stream, Point of Emergence, June, 1906

Eastern margin of Atrevida Glacier.

A. TUNNEL FROM WHICH ESKER STREAM EMERGED IN 1909
The crevassing of 1906 is healed (see Plate XXI, B) and the tunnel resembles that of the earlier years.

B. MARGIN OF ATREVIDA GLACIER, WEST OF ESKER STREAM
Jagged ice cliff on right where, in 1905, it was easy to ascend, over moraine waste, up a moderate slope. From this cliff blocks of ice and bowlders were constantly falling while the photograph was being taken. Note the trees being buried in the morainic débris and overturned by the ice shove. Photograph taken July 10, 1906.

Plate XXIII

Margin of Atrevida Glacier
Against the forest into which it was advancing in 1906 (see Plate XXII, B). Photograph, 1909.

A. Moraine Covered Ice of Atrevida Glacier
In contact with forest trees as a result of the 1905-6 advance. Photograph, 1909.

B. The White Ice of the Future Interior Flat
This appeared in the Atrevida Glacier bulb during the advance of 1905–6. Photograph, 1909.

Plate XXV

A.

B.

Western Margin of the Bulb of Atrevida Glacier
In 1906 (upper picture) and 1909 (lower picture) from Station B (Map 2), on Terrace Point.

PLATE XXVI

ATREVIDA GLACIER FROM AMPHITHEATRE KNOB

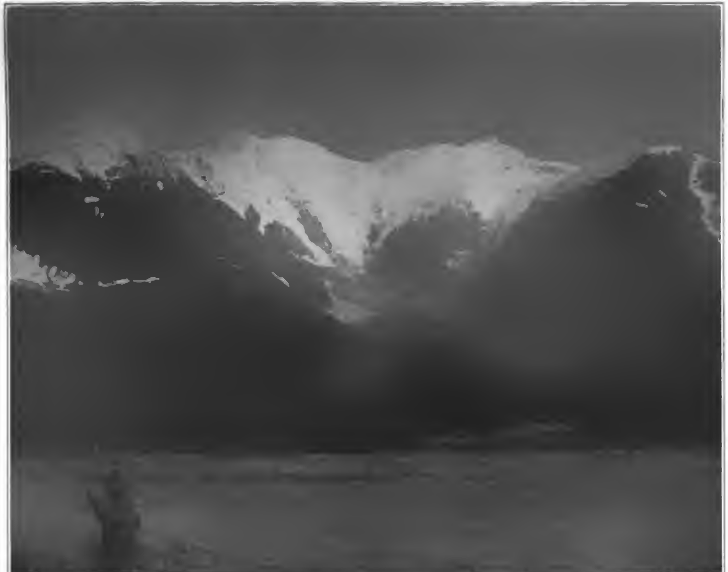
The crevassed bulb of Atrevida Glacier with Floral Hills in background. Morainic waste of Lucia Glacier between Atrevi
Glacier and the Floral Hills. Photograph from Station A (Map 2), on June 29, 1906. See Plate XXVII.

For description, see Plate XXVI. The two pictures form a continuous panorama.

Plate XXVIII

Galiano Glacier in 1909

PLATE XXIX



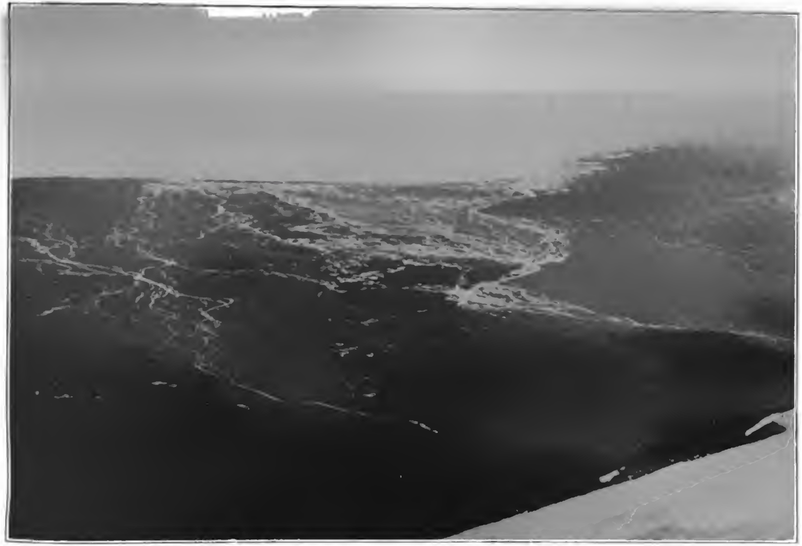
GALIANO GLACIER WITH ITS WOODED PIEDMONT BULB

This vegetation as well as the outwash gravel plain in the foreground were destroyed before 1905. Photograph by I. C. Russell, 1890.

A. Morainic Hills near Galiano Glacier

B. Torn and Broken Stumps of Mature Alder
On the face of Galiano Glacier bulb, where they were growing at least as late as 1890, but were destroyed before 1905.

PLATE XXXI



LOWER PART OF ESKER STREAM ALLUVIAL FAN

Where the distributaries are spread out among the morainic hummocks of former expanded Galiano Glacier. View from crest of Amphitheatre Knob, Station A (Map 2), at elevation of 2522 feet, June 29, 1906.

Plate XXXII

A. Vegetation on Piedmont Bulb of Galiano Glacier in 1909

B. Marginal Stream of Galiano Glacier in 1909

in 1909. Well within the mountain valley the marginal valley disappears, the glacier merging into talus from the mountain, and still higher into snow fans where snow has avalanched down the mountain sides in such masses that summer melting at that elevation is incapable of removing it. The marginal valleys and the marginal streams present some interesting features which are described in later pages.

Beyond the mountains the Galiano Glacier expands into a piedmont bulb more than double the width of the valley glacier. It does not expand westward to a great extent, chiefly because the mountain wall extends farther out in that direction, but partly because it is trimmed on that side by the large glacial stream that flows down the western marginal valley and then turns eastward to enter Yakutat Bay. This expanded bulb is completely covered with ablation moraine, and in the outer portion ice can be seen in only a few small patches; in fact in the extreme eastern part we were unable to find any ice, though its presence was confidently inferred from the height of the moraine, its irregular surface, and the great amount of cold water that emerges from it. The margin of the ice bulb is an irregular, hummocky embankment whose base is essentially at sea level, and whose eastern end lies just back of the beach. The surface of the ice bulb is a maze of hummocks and depressions with the roughly crescentic arrangement so typical of ablation moraines on expanded piedmont bulbs. That this glacier had the same general condition in 1890 as in 1909 is proved both by Russell's photographs and by his description. He says "Débris-covered ice streams, too small to reach the water, are typical of a large class of glaciers in southern Alaska, which are slowly wasting away and have become buried beneath débris concentrated at the surface by reason of their own melting. The Galiano Glacier is a good example of this class." [1]

Along the southern base of the glacier bulb an alluvial fan is being built up by the stream from the western margin of the Galiano and to a lesser degree by small streams that issue from the glacier front. Beyond this, during the period of our observation, was a series of low morainic hummocks between which flow the distributaries of Esker Stream from the Atrevida Glacier. There is good reason for believing that ice still exists beneath this entire area and that, in reality, what seems to be the terminus of the Galiano Glacier bulb, as described above, is only the margin of a part of the glacier which remains higher than the rest because of a more recent supply of ice and of its protection from ablation by the morainic cover. The depressed area of the piedmont bulb beyond the apparent front of the glacier is interpreted, in part at least, as an interior flat, lowered by melting until it became the seat of alluvial fan deposit; but the area beyond this portion, which consists of morainic hummocks, probably represents the moraine-covered outermost portion of the piedmont bulb. The Galiano Glacier is believed to expand beyond the mountain front into a bulb with many times the area of that which is now visible, and to reach five miles or more from the mountain base, attaining a width of two or three miles in its broadest portion. That the bulb is not wider is due to the fact that it forms the shore of this part of Yakutat Bay, and has, therefore, been cut off on the eastern side.

This interpretation of conditions in front of the visible piedmont bulb of the Galiano Glacier is based upon a number of facts, the most significant of which are a series of notable changes in condition between 1890 and 1905, and between 1905 and 1909. These changes affected both the visible piedmont area and the region to the south of it. Their

[1] Russell I. C., An Expedition to Mount St. Elias, Alaska, Nat. Geog. Mag., Vol. III, 1891, p. 101.

nature and significance can best be stated by showing the condition of the area in the several periods of observation, namely in 1890 and 1891 by Russell, in 1905 and 1906 by the U. S. Geological Survey parties, and in 1909 and 1910 by the National Geographic Society expedition.

Conditions in 1890–1891. In his expedition of 1890 Professor Russell landed on the west side of Yakutat Bay near Galiano Glacier, and he has given us both a description of it and several photographs in 1890 and 1891 from which we can determine its principal characteristics. He states,[1] for example, that "a little way back from the shore clumps of alders, interspersed with spruce trees, marked the beginning of the forest which covered the hills toward the west and southwest." Near this rose a densely wooded hill "about 300 feet high, with a curving outline, convex southward," which lay "at the mouth of a steep gorge in the hills," plainly referring to the Galiano valley. Attracted by the peculiar appearance of the hill he forced his way "through the dense thickets" to its top where he "found a large kettle-shaped depression, the sides of which were solid walls of ice 50 feet high." Continuing toward the "steep gorge in the hills" he forced his "way for nearly a mile through dense thickets" before reaching the "broad open fields of rock and dirt" which "completely conceal the ice and form a barren, rugged surface, the picture of desolation."

This description does not fit any conditions existing in this vicinity in 1905 or later years, and we might suspect that some other region was referred to were it not for the fact that Russell's photographs show clearly that the outer Galiano Glacier at that time was covered with vegetation. In his photographs the entire visible bulb beyond the mountains is darkened by vegetation (Pl. XXIX). From these photographs it is evident that the continuous vegetation extended well within the mountain valley, and on the inner side became thinner, and was then succeeded by isolated patches, and finally by the moraine desert. This vegetation on Galiano Glacier bulb is shown in a Boundary Survey photograph in 1895, also proving that the vegetation was present four years before the 1899 earthquakes. The same photograph shows vegetation out beyond the present bulb of Galiano Glacier where now there is only a barren alluvial fan.

Russell's photographs also show an even-surfaced alluvial fan continuous in 1890 and 1891 from the visible margin of the piedmont bulb southward for two miles or more. Over this fan the waters of Esker Stream and the stream from the western margin of Galiano Glacier flowed in many distributaries. The surface of the fan was diversified only by the stream channels and was so smooth that the streams were evidently shifting constantly, for no vegetation, not even annual plants, grew over most of the surface. It was such a perfect alluvial fan that one would never suspect that it was built on glacier ice. Two small, low patches of alder and two low, narrow, alder-covered ridges rose above a portion of the alluvial fan near the sea, proving that these sections rose high enough to be above the reach of the shifting streams; but there is no indication as to their nature. Probably, however, they represent moraine knolls and ridges not yet consumed or buried by the fan-building streams. One of Russell's photographs was taken from a gravel knoll which remained in 1905 and 1909. It is an overridden gravel remnant, bearing bowlders and till on its surface, showing that after its formation it was covered by glacier ice.

Conditions in 1905. When we visited this region in 1905 it required only a glance to see that there had been important changes in the interval of 15 years since Russell's

[1] Russell, I. C., An Expedition to Mount St. Elias, Alaska, Nat. Geog. Mag., Vol. III, 1891, pp. 86–89.

observations, and in the ten years since the Boundary Survey photograph was made. Accordingly, although the full significance of the changes were not then clear to us, we made comparative observations and photographs from Russell's sites and studied the glacier and vicinity with his descriptions before us. There were two noteworthy changes, one in the glacier itself, the other in the alluvial fan just described.

On the glacier the alder and forest cover was entirely destroyed, and our 1905 photographs show barren ablation moraine from far up the mountain valley clear down to the visible front of the piedmont bulb. Even on the most stagnant eastern portion the forest was gone, and here were found quantities of dead wood, including mature tree trunks. In other places mature alder littered the surface of the piedmont bulb or was mixed with the morainic débris, and in two places the dead trunks were seen in considerable numbers, upright and in place. Near the centre of the glacier, 60 feet above sea level, we found twenty or more small trees, the oldest with 25 annual rings (Pl. XXX, B). They increased in size upward, as if standing upside down, but had downward-pointing roots for the lower visible 3 or 4 feet. These roots extended from shaggy bark, proving that the upward-expanding tree remnants were not tap roots, and indicating that the last months of life of these trees involved such adverse circumstances of soil encroachment that roots were sent out at higher and higher levels as the lower portions were buried. A yard or less above the highest roots the trees were torn off and the frayed splinters were bent down.

The alluvial fan was as completely altered as the forest on the glacier. Not only could it no longer be seen from the sites of Russell's pictures, which we carefully located, but there was no such perfect fan anywhere between the Galiano Glacier and the Kwik River. In its place was a series of low, undulating moraine hills (Pl. XXX, A), some of them strewn with angular bowlders, such as characterize ablation moraine, between which flowed the muddy branches of Esker and Galiano streams. These streams were building up the depressions in the moraine and here and there were trimming the edges of the knolls. Besides the muddy glacial streams there were innumerable cold, clear-water streams emerging from the moraine, and small pools of clear, cold water in the depressions. There was a roughly concentric arrangement of the morainic hills with the concave side toward the Galiano Glacier. These morainic hillocks extended down to the shore of Yakutat Bay (Pl. XXXI); and that they also extended out under its waters was indicated by the pronounced shallowness of the water off shore along this coast, which caused icebergs to be stranded in far larger numbers, and to a greater distance out from the coast than is common along the shores of the bay. The moraine hillocks were found not only in the area of the former alluvial fan, but extended even to a distance of two or three miles to the south of Russell's photographic site. In this outer portion we have no photographic proof of the condition in 1890, but the close resemblance of the low, barren, conical hills, ridges and groups of morainic hills in this area to those already described leads us to class them together as to cause. This conclusion is further supported by the presence here of great numbers of tree trunks which littered the surface. Some of these, especially in the depressions near the coast, were doubtless brought and perhaps destroyed by the earthquake water wave which swept this coast in September, 1899; but others, resting above the line of driftwood that this wave brought, and even embedded in the moraine of the hillocks, cannot have been brought by the earthquake wave.

We feel confident, therefore, that some great change had occurred here which had de-

stroyed the existing forest and roughened the surface by the introduction of morainic knolls since 1890, and that this change had operated not only on the visible piedmont bulb of Galiano Glacier, but also for a distance of four or five miles south of where ice is now visible.

From our study of the photographs of 1890 in comparison with the conditions in 1905 and later years we believe that there has been an advance of the visible front of the piedmont bulb of Galiano Glacier, and a recession of the coast line immediately to the southeast of it; but the photographs were not taken from the right points of view to make it possible to state this as a positive fact.

While there had been such widespread and complete destruction of alder and other growth of vegetation over a wide area, it is a noteworthy fact that the vegetation had again begun to grow on this area in 1905. On the visible piedmont bulb were many individuals and some clusters of young alder, five or six years old, while similar growth, with an abundance of annual and perennial plants, had taken possession of favorable portions of the moraine hillocks that have replaced the alluvial fan. If additional proof were necessary of the recency of the appearance of this morainic surface the condition of the vegetation would furnish it. In this climate, close by the sea, alder, cottonwood and spruce thrive: in fact, they extend even farther up the fiord and also clothe the hill slopes that rise to the west of this area. The morainic soil is unquestionably capable of supporting such growth, for in the interval between 1905 and 1909 the development of the alder has been remarkable. That such vegetation does not already thickly clothe the moraine areas can be accounted for only as a result of the fact, proved by other evidence equally, that the moraine has not been here long enough.

Interpretation of the Changes. These phenomena, which puzzled us greatly in 1905,[1] were interpreted in 1906 by the senior author as follows.[2] Galiano Glacier possessed features in 1905 not shared by the Black Glacier on the east and the Atrevida Glacier on the west. The puzzling new morainic hillocks and ridges were concentric and faced Galiano Glacier whose forest had been destroyed since 1890. This destruction followed a long period of stagnation, as shown by the dense alder thicket and forest that grew on the glacier, where we found large broken trunks in 1905, one of which had 75 annual rings. On the border of the alluvial fan, also, quiet conditions had permitted the growth of trees for at least 60 years, and for a long time alluviation had been in progress, building the perfect alluvial fan now destroyed. The abrupt change took place sometime between 1890 and 1905, destroyed the forest on the glacier and for 4 or 5 miles beyond it in the alluvial flat, up through which were pushed a series of hummocks and moraine patches on which vegetation was in 1905 just beginning to grow again. That the change occurred later than 1895 is proved by the Boundary Survey photograph of that year; and that it was not earlier than 1899 is indicated by the age of the new alder plants on the glacier and on the outer moraine hillocks.

To account for this change by mere slumping would involve some special explanation of the 60 or 75 years of immunity from slumping necessitated by the forest and the sudden introduction of excessive slumping over an area of several square miles, and would not explain the broken and splintered tree stubs. Shattering of buried ice by earthquake shaking would share the last disadvantage; furthermore, it would be most remarkable

[1] Tarr, R. S. and Martin, Lawrence, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 152–3.

[2] Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 73–74.

to have had shattering in the broad area of the lower Galiano without corresponding breaking in the neighboring Atrevida Glacier. The natives would have known it if there had been a five-mile advance. The only remaining hypothesis is that the Galiano Glacier extends out here for 4 or 5 miles as a broad, lowered, moraine-covered piedmont bulb, the other portion being reduced by ablation, which has automatically checked melting, and that this part of the glacier had been nearly buried by alluvial fan deposits before 1890, forests growing upon the unburied knobs of moraine and upon higher parts of the fan. Later, when Galiano Glacier moved forward, under the impulse derived from avalanching during the 1899 earthquakes, the forest on the lower end of the visible bulb was destroyed and the thrust even caused the buried ice to push up through the alluvial fan 4 or 5 miles away, destroying the forest there, causing more rapid melting, and giving rise to the distribution of alluvial fan and moraine patches, with new vegetation, seen in 1905.

Observation here and elsewhere in 1906 and 1909 give to this interpretation a probability which we could hardly dare assign to it in 1905 when it stood as the only known case of such a change due to earthquake shaking. Now the main facts observed here are duplicated in other cases where the facts are matters of direct observations, not of interpretation alone. The only serious difficulties that occur to us in accepting this explanation are the remarkably quick response of this glacier and the great extent of piedmont area affected, proportionately more than in other cases. The phenomena seem incapable of any other explanation, however, and we are forced to the conclusion that for some reason this small glacier responded not only quickly, but with such vigor as to affect a great area of stagnant, buried ice. There may, of course, be a very deep ice bulb buried here through which the impulse was communicated with such vigor as to spread far out toward its periphery.

Naturally we have searched carefully in the hope of finding the ice itself in this outer disturbed area; but our search has not been successful. Aside from the changes recorded, however, there is some evidence that the ice still lies beneath the deposits beyond the visible ice bulb. Perhaps the most noteworthy evidence is the great abundance of clear, cold water that oozes from all parts of the newly-elevated morainic area. There is no known source for this water, which far exceeds the rainfall, other than buried ice, and we have uniformly found the same condition of abundant cool springs and pools where we have other reasons to suspect the presence of buried ice; but we have not found this condition in localities where buried ice is not suspected on the basis of independent evidence. There are also a few areas where slumping has recently occurred, and there are some cracks in the soil, and numerous boggy places, all of which are common phenomena in places where moraine rests on buried ice. From these evidences alone we would have confidently inferred the existence of buried ice here, even though we had not been forced to this same conclusion by the striking changes between 1890 and 1905.

Granting the existence of a buried, stagnant, ice block here we are led to inquire how it came to be in the condition observed in 1890. First it was doubtless an expanded piedmont bulb, and such a condition it would maintain so long as supply equalled general ablation; but when ablation exceeded supply it would begin to waste away and ablation moraine would cover its surface more and more deeply. The further wastage would then be differential and those parts which had least moraine would be lowered most rapidly. It is a common condition in piedmont ice bulbs to find an area immediately in front of

the mountains on which, for some reason, little débris exists. This is typically illustrated in the Variegated Glacier in this region and in the Allen Glacier in the Copper River region. Such a clear ice area developed in Atrevida Glacier in 1906, as already stated. We infer the existence of a similar clear ice area in the Galiano Glacier, which by more rapid ablation was reduced to the condition first of an interior flat and then, when low enough, to a region of alluviation. It was in the latter condition in 1890, and was then receiving deposits not only from Galiano Stream, but also from the drainage of Atrevida Glacier.

The large ice block which we believe to be buried here must in the course of time receive a thick cover of outwash gravels which have a tendency to preserve it from wastage. But inevitably, in the course of time, if stagnation continues, this ice will completely melt away. If alluviation continues no depression may result, but if glacial streams cease to bring sediment, a large kettle would ultimately mark the site of the buried ice or, since the ocean is so near, an arm of the sea. The present condition and possible future of this region have more than local interest. It is one of many instances of more or less completely-buried masses of stagnant ice in the Alaskan region, which show us in actual existence what was undoubtedly common along the margin of the waning continental ice masses of the Glacial Period. The literature of glacial phenomena does not assign to buried masses of stagnant ice the full importance which they probably had in shaping the topography of marginal deposits. The conditions near Galiano Glacier, and elsewhere in the Yakutat Bay region, give facts which ought to be of value in interpretation of these phenomena.

Condition in 1909. The interval between the visits of 1905 and 1906 was too brief for any notable change in the condition of Galiano Glacier and the region to the south of it. But by 1909 there had been so much change that it was noticeable, and this change was of several kinds. The area of the ablation moraine in the mountain valley had extended farther toward the valley head than in 1905, when it evidently did not extend so far as in 1890. Alder had again spread over a large part of the piedmont area (Pl. XXXII, A) and even into the mountain valley. In most places the alder was young and in scattered individuals and clusters; but in the more stable parts of the piedmont area, and especially on the eastern side, there were already alder thickets through which it was difficult to travel, and in which the bushes were eight or nine years old. There was still much slumping of moraine in the piedmont area, probably because crevasses still existed there, and there were large pits barren of vegetation, but surrounded by alder growth; and there were other places where recent subsidence had overturned the alder that had begun to grow. Evidently the glacier had not yet settled down to a state of stability equal to that which permitted the growth of such a uniform cover of alder as was present in 1890.

There were also notable changes in the marginal valleys on both the east and west sides. That on the east side was much broader than in 1905 and headed farther up in the mountain valley. Only a small stream emerged from the glacier on this side, as was true also in 1905; but it was cutting the glacier back, forming a steep ice cliff which was extending down-stream, and thus was broadening its valley. At its lower end the ice cliff merged into a moraine-covered ice embankment on which young alders were growing. At this point, where the ice cliff was being extended down-stream, the moraine and its alder growth were sliding down the slope and here one saw withered, present-

day 5 to 10 year old alder bushes, some of them still green, and the battered stumps of the earlier alder growth, many of them over 25 years old, all mixed together in the moraine that was sliding off of the glacier margin.

On the west side, also, the marginal stream was extending its valley up into the mountains (Pl. XXXII, B), baring the ice of its load of marginal débris and, in places, developing a steep ice cliff, over a hundred feet high, down which we could not descend, and forcing us up-stream a quarter of a mile to a more gently-sloping portion from which the moraine had not yet been removed. Here we saw a remarkable exhibition of the way in which the ice margin was being stripped of its moraine cover. While eating luncheon by the bank of the marginal stream, just where the bare ice cliff graded into the moraine-covered portion, we watched several small streams issuing from small crevasses and holes in the ice and descending torrentially over and through the moraine talus at the ice base. All of a sudden a landslide of considerable proportions slid down the ice face where one of the streams was issuing from the moraine (below A, Pl. XXXIII). It was so charged with water that it formed a veritable mud-flow in which stones a ton in weight were floated. The avalanche reached the marginal stream and in a few seconds forced it to one side and over the place where we had just then been seated. Slide after slide followed, and in less than fifteen minutes the marginal stream was pushed over into the fringing alder thicket, fully fifty feet from the place where it formerly flowed. In this brief interval a fan was built whose dimensions were estimated to be 100 by 50 feet, with a depth of 6 feet at its front margin and 25 to 30 feet in the center. An area of ice of several score square yards was bared by these landslides.

On both margins of the glacier near these marginal streams we encountered a band of crevassed ice which we did not see in 1905. It was not fresh breaking, such as is caused by advance, and, moreover, its marginal position, and the absence of crevassing in the glacier center and up the valley precludes this explanation. The phenomenon seems related to the development of the marginal drainage just described. Part of it may be due to slumping through undermining, but most of it is apparently due to the removal of débris from older crevasses, probably formed during the last cycle of advance. It is conceived that prior to 1905 the ice was broken, as the Atrevida Glacier was in 1906. This breaking was soon healed by ablation and by the sliding of moraine into the crevasses so that by 1905 they were hidden. Presumably there are similar crevasses all over the deeply moraine-covered ice, but most of them are now buried from view, though here and there one is seen; and there are long, narrow, linear valleys and ridges, which probably mark the sites of still others. The extension of the marginal drainage up the glacier side, doubtless a redevelopment of marginal drainage toward the condition that existed before the advance, has sharpened the cliff and exposed the ice, and from this the streams emerge heavily charged with débris. It is thought that the removal of this débris is opening up the marginal crevasses. Further down the glacier, where the débris is well stripped off, many crevasses are revealed. Another indication that the explanation proposed is correct is the fact that water is emerging from orifices part way up the ice cliff, as if from the bottoms of crevasses. In more mature ice drainage the water would find its way to the ice bottom through moulins and emerge from the ice base rather from the middle of the marginal slope.

A third notable change in this region in the interval between our visits of 1905 and 1909 is in the coast line. Just south of the glacier front the coast had been completely altered

in detail, and although in a region of deposit, where the normal change should be an outward growth, it had receded several yards and a gravel knoll that stood there in 1905 had been nearly consumed. The site of our camp of 1905, which was on the beach at the base of this knoll had disappeared also. Slumping of the alluvial fan back of the beach was also indicated by the fact that, although this area had for four years received the drainage of the large stream that flowed from the western margin of Galiano Glacier, its bottom had not yet been raised above the reach of the tide. Indeed, it seemed even lower than it was in 1905.

Out on the alluvial fan-moraine area in front of the visible ice bulb of the Galiano there had been much change in detail, but principally of two kinds. In the first place there had been a great increase in the amount and density of vegetation. On the higher and more extensive morainic knolls there were now alder thickets through which in places one found it difficult to travel, whereas in 1905 there were no such areas. On the lower knolls there was less noticeable increase in vegetation, but we had good evidence of increase in one place when we attempted to find the site of one of our 1905 photographs and, in looking for the foreground stones, found them partly hidden by alders, whereas in 1905 there were only annual and perennial plants and a few scattered alders one or two years old. The entire foreground (compare vegetation near stones marked A, B, C, and D in Pl. XXXIV) was completely altered by growth of vegetation.

A second noteworthy change is that of deposit among the morainic hillocks. The process of filling the depressions, and thus of submerging the hillocks beneath gravel deposit, had proceeded so far as to be noticeable in some places; but we had no data for determining whether at the same time there had been a lowering of the hillocks by the melting out of the underlying ice, though this must inevitably have been the case, to some extent at least, for large quantities of cold water still issued from the moraine and this had without doubt been continuing during the intervening summers. The water that escaped represented ice that was lost by melting, and this loss must have caused a lowering of the moraine.

It is natural that the anastomosing glacial streams which are building an outwash gravel plain between the moraines should change their courses as they build one portion higher than another. A number of such changes had occurred, but one was specially noteworthy. In 1905 and 1906 the Galiano stream flowed over an alluvial fan which it had built between the visible glacier front and the first range of low morainic hills a quarter of a mile to the south of it. Over this fan the stream wandered in branching and ever-shifting courses, and at times a branch swung over against the moraine which it had trimmed back into a fairly steeply-sloping bank. On the morning of July 28, 1909, the Galiano stream was flowing as usual, with three large distributaries and several smaller ones, all of which were easily forded, though the northern one was fairly deep and swift; but a small branch had found its way southward through a gap in the moraine. When we returned at night our progress was barred by a single torrent which outflowed through the gap, and which was crossed with difficulty. All the other stream courses were abandoned, and for the time being the entire lower portion of the alluvial fan received no water or sediment. The new course proved so favorable that it was maintained as long as we remained there (until August 3), and that it was likely to be followed for some time was indicated by the fact that the torrent had entrenched itself in a newly-formed gravel gorge which was extending up-stream.

West of the moraine which had been pushed up through the alluvial fan of 1890 is a very perfect fan built by Esker Stream. Over this the glacial waters flow in many branches which are constantly shifting (Pl. XXXI). It is the most perfect alluvial fan in that vicinity, but as a perfect alluvial fan it terminates against the morainic hillocks and ridges of the buried Galiano bulb, and the waters pass down among them, making deposits in the depressions; but in 1890 the upper fan was continuous with the lower one, now destroyed. In each period of observation (1890, 1905, 1906 and 1909) the upper fan has been not only growing higher but extending laterally. The latter condition is brought out with striking clearness by reason of the fact that the alluvial fan was bordered on the north by a forest of cottonwood and spruce trees into which the glacial torrent is pouring its flood of water and sediment, burying and killing many of the trees (Pl. XXXV, A). Some, only slightly buried, are still growing in the barren alluvial fan; others stand there dead, and one of our packers who is familiar with the Alaskan forest, judging from the position and nature of the limbs, estimated that the gravel rose up on several of them to a height of twenty-five feet. Back in the forest were gravel terraces on which a mature forest grew, and in which, beneath the living forest, were dead mature trunks in place, proving a still older period of burial at a higher level. It is believed that these deposits were laid down when the expanded bulb of Galiano Glacier stood up as a barrier across Esker Stream, forcing the deposit of its sediment load at a higher level than at present. Thus the region has been the theatre of extensive past as well as present day changes.

The Cycle of Change. Galiano Glacier is the only instance in the Yakutat Bay region in which a practically completed cycle of change resulting from earthquake advance has been observed. It seems well, therefore, to briefly summarize the succession of events as observed and inferred.

It is inferred that at some period, long before 1890, Galiano Glacier flowed out of its mountain valley and expanded into a huge bulb extending some five miles or more from the mountains, expanding somewhat toward the west, and toward the east entering the waters of Yakutat Bay. Just beyond the mountain front was an area of clear ice, but beyond it the piedmont ice became covered with ablation moraine. Ablation proceeded to lower the surface, most rapidly in the area of clear ice, but finally reduced the entire outer area nearly or quite to sea level. A glacial stream from Atrevida Glacier, Esker Stream, built an alluvial fan up to the western side of this piedmont ice area. In the meantime a stream, developed on the western side of the Galiano Glacier, found escape into the interior flat that had developed in the clear ice area and began to deposit there; and in this work of deposit it was ultimately joined by Esker Stream, which hitherto had been deflected southward by the expanded Galiano ice bulb. Meanwhile forest had developed on the lowered moraine to the south of the alluvial fan of the interior flat; and a growth of alder, and, toward the east, of cottonwood and spruce trees, had covered the ice bulb that still rose above the interior flat and between it and the mountain. The glacier was in this condition when Russell studied it in 1890, and when the Boundary surveyors photographed it in 1895, and doubtless continued in this condition until 1899 or 1900.

Then came the series of earthquakes in September, 1899, which shook down into the upper portion of the Galiano valley a great series of avalanches of snow and ice, the scars of some of which were easily recognizable in 1905 by comparison with 1890 photographs. Quickly, and presumably as early as 1900, judging by the age of the new alders found in

1905 and by the extent of healing of the glacier in that year, the entire Galiano Glacier from its valley head to the periphery of its buried piedmont bulb was under the influence of a spasmodic advance, which broke the surface of the glacier and pushed up through the alluvial fan as well as the area of morainic hillocks beyond it, at the same time destroying the alluvial fan, overturning the forest that grew on the glacier and that which was growing to the south of the alluvial fan. Ablation on the deeply moraine-covered ice soon hid the crevasses, partly by melting down of the surface, but largely by the sliding of the moraine into the crevasses, so that in 1905 the glacier surface showed no signs of the pronounced breakage of a few years before, which would be recognized by an observer unfamiliar with the phenomenon. This latter point would perhaps seem improbable were it not for the evidence that the advancing glaciers of 1906 furnish concerning the rapidity with which such a broken surface is healed by ablation.

Now the glacier is returning to its former condition and the evidence of it was especially seen along the sides in 1909 where marginal drainage was again developing, and in doing so revealing some of the crevasses that were formed during the advance and that since then had been hidden by the ablation moraine. It was also shown by the immaturity of the ice drainage. The return of stagnation has again permitted the encroachment of vegetation upon the glacier, and it is rapidly spreading and developing alder thickets; but the presence of buried crevasses beneath the moraine, and the necessity of development of drainage, are causing so much slumping that the advance of the vegetation is interrupted here and there.

Changes were also in progress on the outer area of stagnant, buried ice, but this had not yet proceeded far and the morainic hillocks still rose prominently above the surface. However, vegetation was taking root here also, alluviation was in progress between the hillocks, and probably the surface of the moraines was being lowered by the melting of the buried ice. There was a tendency toward the return of the condition of 1890, but it will be many years before this work is complete, for there were many irregularities to be removed or graded up before an even-surfaced alluvial fan is again built here; and, although glacial streams bore great quantities of sediment, and were relatively rapid agencies of deposition, there was a large area to be covered and much deposit was necessary. At the observed rate of change it is probable that a quarter of a century hence the signs of the spasmodic breaking of the Galiano Glacier will still be visible here, while in the valley portion and its small inner piedmont bulb all signs of the advance will have disappeared. Ablation on the glacier itself is far more rapid than on the outer bulb which bears so thick a coat of morainic débris and alluvial fan gravels that no ice has been observed in it.

In 1910 and 1913 the Galiano Glacier seemed, from the bay, to be essentially as in 1909, except for the increased covering of vegetation.

Black Glacier

Black Glacier, a little over a mile east of Galiano Glacier (Map 2) resembles the Galiano in several respects, though in others the two glaciers are quite different. Although shorter and narrower than the Galiano, the Black Glacier occupies a similar steep-sided, short valley and is supplied with ice by several steeply-descending cascading tributaries at the valley head. Its surface to a point far up in the mountain valley is covered with ablation moraine, and in the lower portion, in the areas of deepest moraine, there is considerable alder growth, forming small thickets. There is no continuous alder thicket,

such as existed on the Galiano in 1890, but some of the bushes are quite mature. A noteworthy difference between the two glaciers is the fact that the lower Black Glacier is a ridge, with a well-defined marginal valley on either side; and a still more notable difference is the absence of a pronounced piedmont bulb at the lower end of the Black Glacier. It barely passes beyond the mountain front and expands slightly at the very lowermost end; but no piedmont bulb is produced.

Although the earthquakes of September, 1899, must have shaken the Black Glacier as much as the Galiano, we have been unable to find any evidence of an advance; and there are several facts which prove that if this glacier did respond at all to the influence of earthquake shaking it was only to a very limited extent. Russell photographed the glacier from its alluvial fan in 1890, and in 1905 we re-occupied the site of this picture and were unable to detect any distinctive change in appearance. That the glacier has not advanced to any extent since 1890 is proved by the presence of a mature cottonwood forest about 250 yards from the glacier front, which also appears in Russell's picture. This photograph (Pl. XXXV, B) was taken at such a distance that we cannot be certain of the space that then separated the glacier from the forest, but it seems to have been about the same in 1890 as in 1905. Instead of advance, Black Glacier front gives evidence of recent recession, for in 1905 there was a low, moraine-covered stagnant ice mass about 100 feet from the glacier front, and between it and the glacier there was no mature alder growth. But just outside this stagnant ice mass, and within a few feet of it, were alder bushes from ten to fifteen years old. The interpretation that we place on these facts is that in 1890 the ice front stood where the stagnant ice now rests, and that since then there has been recession of about 100 feet. Prior to 1890 the ice had been receding over the area of about 200 yards now covered by alder, and for at least a half century had not been farther out than the border of the cottonwood forest.

Confirmatory evidence of the conclusion that Black Glacier has not been subjected to a recent advance as a result of earthquake shaking is supplied by two facts already mentioned, namely, the mature alder growth still standing on the glacier, and the ridge form of the glacier itself. The ridge extends about a mile from the ice front, to an elevation of about 1100 feet, and its surface rises up the valley at a high angle, averaging about 20° in the first half mile. It is a single, narrow, moraine-covered axial ridge, sloping toward the valley walls often at an angle of 30° or 35°. Bare ice shows in many places, and stones frequently roll down the marginal slopes; but the moraine on the ridge top is thicker, 5 or 6 feet, and ice is rarely revealed there. Alders grow in this thick moraine, one that was cut down in 1905 being 13 years old. Some are killed by sliding and slumping of the soil, but this did not seem to be due to ice movement, for there were no crevasses in the central ridge in the lower half mile of the glacier. Farther up the glacier, where crevasses are present, deep morainic soil and alder bushes are not found. Alders 5 or 6 years old grow on the glacier up to 1000 feet, a mile from the end of the glacier. Beyond this the ridge flattens and disappears but in both 1905 and 1909 ablation moraine covered the glacier clear to the head.

Had there been notable recent advance of this glacier the ridge would of necessity have been destroyed and the glacier put into closer contact with the mountain wall. It is inconceivable that so pronounced a ridge could have been formed by the development of marginal valleys in the interval between 1899 and 1905. Also, the mature alders could not have survived an advance and breaking of any considerable amount. Those that are

being overturned are easily accounted for by normal slumping of the morainic soil, accentuated here, perhaps, by the continued development of the ridge form of the glacier through ablation.

The facts observed force us to the conclusion that Black Glacier has experienced no notable advance under the impulse of earthquake shaking. This does not, of course, imply that there has been no change as a result of the earthquake, for in the upper portion there may have been an advance and breaking of such slight proportions that it was not communicated to the lower portion. Possibly the small area of crevassing a little more than a mile from the terminus is a relic of such an advance.

The contrast between Black and Galiano Glacier in this respect is notable, and it is this contrast which to us constitutes the most interesting feature of the Black Glacier. The neighboring Galiano Glacier supplies unquestionable proof of a great advance and breaking, while Black Glacier furnishes equally convincing proof of the absence of such advance and breaking; yet both receive their snow supplies from the same earthquake-shaken area. No other reason for this difference appears than the difference in size of the glaciers; and, of course, the related difference in extent of supply area. It must be concluded that there was not a sufficient amount of unstable snow and ice available in the supply ground of Black Glacier to give the impulse necessary for advance and breakage down to the glacier terminus. It may also be true that the narrow, and perhaps thin, ridge of ice that forms the lower Black Glacier was not a large enough mass of ice to transmit the thrust. This could only be true, though, if the thrust itself was weak; for a strong thrust would have broken and pushed forward even a rigid ice body that stood in the way. In 1905 it seemed to us a possibility, though not very probable, that the impulse of earthquake advance in the case of this glacier has for some reason been retarded and that its effects would be noted in later years. We accordingly examined it again in 1906 and in 1909, but could see no change. In 1910 and 1913 there was no change that was visible from Yakutat Bay. That an advance could be longer delayed seems inconceivable in a glacier of such small size, and in view of the notable advance of so many other larger glaciers. We therefore conclude that not only has there been no notable advance up to 1913, but that no future advance of the Black Glacier is to be expected as a result of the earthquakes of 1899.

CHAPTER VI

TURNER, HAENKE, AND HUBBARD GLACIERS

Turner Glacier

General Description. On the western side of Disenchantment Bay, a few miles up the fiord is the Turner Glacier (Pl. XXXVI), the first of the three tidal glaciers that discharges icebergs into Yakutat Bay. It was first described and photographed by Russell in 1890 and 1891; it was photographed by Brabazon of the Canadian Boundary Survey in 1895 and by Bryant in 1897; it was again photographed and described by Gilbert in 1899, and photographed by the U. S. Fish Commission in 1901; and it was studied and photographed by us in 1905, 1906, 1909, 1910 and 1913. We, therefore, have a more complete photographic record of the Turner Glacier than of any other in the Yakutat Bay region.

Russell saw Turner Glacier from Haenke Island[1] on July 3, 1890, and September 5, 1891, and described it in this language:

"From a wild cliff-enclosed valley toward the north, guarded by towering pinnacles and massive cliffs, flows a great glacier, the fountains of which are far back in the heart of the mountains beyond the reach of vision. Having vainly sought an Indian name for this ice-stream, I concluded to christen it the *Dalton glacier*, in honor of John Dalton, a miner and frontiersman now living at Yakutat, who is justly considered the pioneer explorer of the region.[2] The glacier is greatly shattered and pinnacled in descending its steep channel, and on reaching the sea it expands into a broad ice-foot. The last steep descent is made just before gaining the water, and is marked by crevasses and pinnacles of magnificent proportion and beautiful color. This is one of the few glaciers in the St. Elias region that has well-defined medial and lateral moraines. At the bases of the cliffs on the western side there is a broad, lateral moraine, and in the center, looking like a winding road leading up the glacier, runs a triple-banded ribbon of débris, forming a typical medial moraine. The morainal material carried by the glacier is at last deposited at its foot, or floated away by icebergs, and scattered far and wide over the bottom of Yakutat Bay.

"The glacier expands on entering the water, as is the habit of all glaciers when unconfined and ends in magnificent ice-cliffs some two miles in length."

Gilbert also described Turner Glacier fully in 1899,[3] comparing the conditions with those in 1891, 1895 and 1901.

The description of this glacier published in our reports of the 1905 and 1906 expeditions

[1] Russell, I. C., Nat. Geog. Mag., Vol. III, 1891, pp. 98–99; 13th Ann. Rept. U. S. Geol. Survey, 1894. Pl. XX, facing p. 86.

[2] The authors of this volume agree with this characterization, and regret the subsequent change in the name of this glacier from Dalton to Turner.

[3] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 66–69.

summarizes the main results of the earlier observations down to the close of 1906, and has been freely paraphrased here,[1] with minor additions. Turner Glacier flows from an unknown source upon the slopes of Mt. Cook through a valley nearly a mile wide and with walls five to six thousand feet high (Pl. XXXVII). Its visible slope is moderate and fairly regular, and the rapidly-moving ice is severely crevassed. Where the glacier passes from its mountain valley into Disenchantment Bay it expands to two miles in width (Map 3) and the slope is steepened from less than ten to as much as twenty-five degrees, a change which Gilbert correctly interpreted as indicating the hanging valley condition, due to the overdeepening of the main fiord by the larger, expanded Hubbard Glacier at the time when the smaller Turner Glacier was its tributary. The floor of this valley probably hangs well above the fiord bottom, and perhaps lies even above present water surface. Half a mile back from the terminus, and about half way between the medial and south lateral moraine, there was, in 1906, a massively-crevassed bulge, or dome, suggesting an unexposed nunatak and indicating that the ice was not very thick. Below the steep slope the ice expands into a flat-topped fan or bulb terminated by a crevassed cliff 200 to 220 feet or more high, whose height is shown specifically in Pl. XXXVIII, where it is compared with a New York office building, drawn carefully to the same scale. The levelness of this ice foot suggests that it is either afloat, as Gilbert thought, or else resting on a flat rock surface. If afloat Gilbert estimates that there must be 1500 to 1600 feet of water here, but soundings are not available for determining this. It does not seem probable, however, that the glacier terminus is actually afloat, for the only known depths nearby are 264 feet just west of Haenke Island, 144 to 282 feet between Turner and Hubbard Glaciers near Osier Island, and 720 feet (no bottom) southwest of Haenke Island. The nearest of these is nearly two miles from Turner Glacier, but the depth of 1500 to 1600 feet is not thought probable, for the greatest depths in all Yakutat Bay are only 1002 to 1119 feet (Pl. XCI), and the depth on the side of the bay near the Turner Glacier would normally be considerably less than the maximum. Still another reason for not believing this glacier front to be afloat is the nature of the iceberg discharge. If it were afloat there should be occasional larger masses breaking from the front; but as a matter of fact the icebergs discharged are all of small, or moderate size such as come from the grounded front of a tidal glacier.

The front of Turner Glacier is not a completely-symmetrical bulb or fan, such as forms where glaciers emerge from confining valleys into broader spaces on the land, for iceberg formation and melting in the fiord check complete development of the bulb form. This expanded bulb has, therefore, a truncated terminus with a fairly straight, tidal ice cliff, over two miles long in 1905, flanked by wing-like, moraine-covered points which ended on the land about 2½ miles apart. Icebergs are steadily discharged from the cliff and float seaward with those from Hubbard Glacier, making an iceberg barrier so hard to penetrate that we found it difficult to work a boat up to within a half mile of the south edge of the glacier in 1905, and quite impossible in a half day's work in 1906.

The lateral moraines of Turner Glacier terminated in the tips of the wing-like edges of the ice front, the one on the south being compound. North of it were two narrow, indistinct, parallel, lateral moraines, then thin, narrow bands of débris, and then a

[1] Tarr, R. S. and Martin, Lawrence, Glaciers and Glaciation of Yakutat Bay, Alaska, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, p. 149.

Tarr, R. S., The Yakutat Bay Region, Alaska, Prof. Paper 64, U. S. Geol. Survey, 1909, pp. 39–40.

pronounced medial moraine. Clear ice exceeded débris-laden ice in the south half of the glacier, while in the north half there were no medial moraines and the north lateral moraine was not wide.

The ice cliff is in general regular, reaching its maximum height within a quarter of a mile of the south side. In detail it is made up of innumerable points and recesses, the foremost cape being south of the middle of the glacier. Rapid movement, of undetermined rate, is indicated by severe crevassing and incessant iceberg discharge, even the débris-laden wing-tips being much broken in 1905.

Changes Between 1891 and 1905. Photographs taken in 1905 on Haenke Island from the exact site of Gilbert's 1899 picture, and from as near the site of Russell's 1891 picture as the growth of alder would permit its location, show the following changes, which have previously been announced.[1]

Between 1891 and 1905 Turner Glacier receded, the moraine-veneered tip, and the edge of the clear ice on the south, going back two to three hundred yards and the north edge an equal amount. The centre of the glacier receded a quarter of a mile, as shown by the truncation of the medial moraine. There were minor differences in the distribution of moraine on the ice.

Between 1899 and 1905 the middle of the glacier receded, showing that fully two-thirds of the recession of Turner Glacier between 1891 and 1905 took place between 1899 and 1905. But the north and south glacier tips advanced, both the clear and the débris-laden ice on the north side extending fully a quarter of a mile farther in 1905 than in 1899, while the south side advanced one or two hundred yards and was greatly broken clear up to the edge.

Therefore, as Gilbert showed, there was recession between 1891 and 1899; but there was net recession between 1891 and 1905 with advance of the middle between 1899 and 1901 and with advance and breaking on the edges and retreat in the center between 1901 and 1905.

This general recession is evidenced along the south margin of Turner Glacier where there was a low moraine in 1905 with no ice in it and no vegetation upon its surface. This was perhaps the moraine that the glacier margin touched at the time of Russell's 1891 picture. Just beyond this, and not far above the present glacier surface, the growth of good-sized alders, estimated to be not less than 20 years old, proved that the glacier had not recently been much higher than in 1905. No spruces had advanced within several miles of Turner Glacier.

Condition in 1906 and 1909. The only noticeable modification between 1905 and 1906 was a slight change in the position of the medial moraine, suggesting a minor advance in the central portion of the glacier. In 1909, however, notable change was seen. By comparing the 1906 and 1909 views of Turner Glacier from Osier Island, recession in the northern wing was indicated by the noticeable increase in the area covered by moraine. A visit to the southern margin was made and photographic sites of 1905 and 1906 were re-occupied. Here it was found that the southern wing had receded slightly (Pl. XXXIX) and its crevassed condition had disappeared, being replaced by a moraine-covered, uncrevassed margin, and that the ice over the steep part of the glacier was

[1] Tarr, R. S. and Martin, Lawrence, Glaciers and Glaciation of Yakutat Bay, Alaska, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, p. 154; Tarr, R. S., the Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 40–41.

thinner. This retreat is also shown by comparing 1906 and 1909 photographs from Gilbert's 1000 foot station above Osier Island. Exactly the opposite condition was observed in the tidal portion of the glacier, for in 1909 the point of the ice cliff projected so far as to hide the western margin of Hubbard Glacier which was visible from this photographic site in 1905, indicating an advance of several hundred yards. In 1906 the black point of Haenke Glacier was visible from one of our photographic sites on the south side of Turner Glacier, but in 1909 it had disappeared. This, however, is due in part to the recession of Haenke Glacier.

Condition in 1910 and 1913. There was no great change in Turner Glacier from 1909 to 1910. From Osier Island (Photo sta. C, Map 3) neither the north nor south edges showed any change (Pl. XL, A) but from Gilbert's photographic site 1000 feet above Osier Island (Station D, Map 3) it was apparent in 1910 that there had been a very slight retreat of the south edge and possibly of the north edge also. From both stations an increase of the area of dirty ice on the north margin was apparent. From the crest of Haenke Island (Photo. station A, Map 3), which was not visited in 1909, it was clear that there had been some retreat of the south edge between 1906 and 1910 and, with the recession of Haenke Glacier, of the north edge also. No change in the middle portion was apparent. In September, 1913, when the junior author visited Yakutat Bay with the International Geological Congress, the Turner Glacier was seen to have continued its retreat. The tips of both the north and south wings of the glacier, especially the latter, were so inactive that they had ceased to discharge icebergs. They were covered with ablation moraine to a greater extent than at any time in the previous eight years. Very few icebergs were being discharged from the middle of the glacier cliff. It is clear, therefore, that the relative inactivity of Turner Glacier was continued at least until September, 1913.

Turner Glacier

<table>
<tr><th>Year</th><th colspan="2">Nature of Change</th><th>Based on observations by</th></tr>
<tr><td>1890–1891</td><td></td><td rowspan="4">Retreat</td><td>Russell</td></tr>
<tr><td>1891–1895</td><td>Retreat</td><td>Boundary Survey</td></tr>
<tr><td>1895–1897</td><td rowspan="2">Retreat</td><td>Bryant</td></tr>
<tr><td>1891–1899</td><td>Gilbert</td></tr>
<tr><td>1899–1901</td><td colspan="2">Advance middle and beginning in edges</td><td>Fish Commission</td></tr>
<tr><td>1901–1905</td><td colspan="2">Advance north and south edges, retreat middle</td><td>Tarr and Martin</td></tr>
<tr><td>1905–1906</td><td colspan="2">Advance of middle, retreat of north and south edges</td><td>Tarr</td></tr>
<tr><td>1906–1909</td><td colspan="2">Retreat of north and south edges</td><td>Tarr and Martin</td></tr>
<tr><td>1909–1910</td><td colspan="2">Retreat</td><td>Martin</td></tr>
<tr><td>1910–1913</td><td colspan="2">Retreat, especially at south edge</td><td>Martin</td></tr>
</table>

The general recession from 1891 to 1913 is thought to be normal, perhaps due to decreasing snow supply, perhaps slightly accentuated immediately after 1899 by the earthquake shaking in that year. It is assumed that Turner Glacier has not yet been subjected to any profound spasmodic advance, such as some of the other glaciers of the region have experienced, but that the two periods of slight forward thrust between 1899 and 1901 and between 1905 and 1906 represent only the response to moderate thrusts from some of the tributaries overloaded through earthquake avalanching. A great spasmodic forward movement may yet be in store for this glacier; or the effect of the earthquake may express itself only by similar repeated small advances.

It is unfortunate that none of the tidal glaciers of Yakutat Bay have so far responded to the earthquake shaking, as some of the glaciers which terminate on the land have done. We cannot, therefore, tell what is to be expected as a result of such an advance. We would infer, however, that when such an advance occurs it will be a notable one, for the tidal glaciers are already moving with such rapidity that they are broken by crevasses from side to side, and we would expect the addition of a vigorous thrust to such rapidly-moving glaciers to push their fronts rapidly forward. Opposing this, however, is the tidal condition of the glaciers, owing to which advance under a forward thrust would be reduced by iceberg formation, especially if a grounded glacier front were pushed far enough forward to float, or if the glacier terminus is already afloat. Another difference between the tidal and non-tidal glaciers worthy of mention in this connection is that an advance in the former could not be detected by crevassing alone, as it might be in a non-tidal glacier, because their surfaces are already crevassed as much as is possible. For evidence of change in such a glacier in response to the earthquake shaking one would have to depend upon actual advance of the front, and upon changes along the margin.

When one of the tidal glaciers does push forward it will be a matter of great interest to observe the nature of the change not only because of the differences in condition, as compared with those advancing glaciers already observed, but also as an indication of the extent to which it is possible for these large, vigorous glaciers to advance by this cause. It may have a distinct bearing on the interpretation of earlier changes of great extent to which the glaciers of this region have been subjected. That these glaciers have not yet advanced need not necessarily be interpreted as indication that they will not do so. They are all long glaciers, with many tributaries, and it is reasonable to expect that a considerable period of time must elapse before they respond to the down-shaking of snow and ice in their remote reservoirs. This is perhaps less true of the Turner than of the Hubbard and Nunatak glaciers, but even the Turner is one of the longest and most vigorous of the glaciers of Yakutat Bay; indeed, judging from the discharge of ice-bergs from its front it is more vigorous than the Nunatak Glacier. It is of course possible that neither Turner nor any of the other great glaciers may ever advance notably, not because of failure to receive great additions of snow and ice to their reservoirs by avalanching, for this is inconceivable, but by reason of minor responses, due to thrusts from individual tributaries at different times, dissipating the energy of the thrust that in other glaciers has been concentrated in one great spasmodic advance. While such a condition is a possibility, it is hardly probable and we feel warranted in predicting that in time the Turner and other tidal glaciers of the Yakutat Bay region will in all probability show in a spectacular way the result of such an advance. It is to be hoped that when this does occur, some observer will be on hand to witness its nature and extent.

Haenke and Miller Glaciers

General Description. Immediately north of Turner Glacier, and between it and Hubbard Glacier are two much smaller glaciers, each of which, however, is a great deal larger than Galiano Glacier. The southernmost of these has been named Haenke, the other had no name until now.[1] Both of these glaciers have their sources in numerous small tributaries, some of them cascading, which are fed from the extensive snowfields that cover the lofty mountains to the northwest of the head of Disenchantment Bay. Haenke Glacier is the longer of the two, its grade is also steeper than that of its neighbor, and it is set more deeply in its valley. Both are normal valley glaciers, and until 1906 both extended almost to the ends of their valleys, but failed to reach the fiord, from which they were separated by alluvial fans built by the streams to which their melting gave birth. In each case the strip of alluvial fan was a half mile to a mile wide, measured from the glacier front to the fiord, and the fans proved conclusively that these glaciers had not recently been tidal, for many years must have been required to build such broad, perfect, alluvial fans. Although they appear in the photographs of previous expeditions, these glaciers had not, until 1906, attracted enough attention in this region of much larger, more accessible glaciers, to have been given descriptions or even names.

From the photographs taken prior to 1906 it is perfectly clear that in 1891, 1899 and 1905 these two glaciers were essentially stagnant at their lower ends, which were covered with an extensive waste of black shale ablation moraine, much as Black Glacier still is. Far up the mountain valley, in each case, there was abundant crevassing, and in the upper reaches of each there was evidence of distinct activity; but for two or three miles from the fronts of the glaciers there were no crevasses that could be seen from the distance at which these glaciers were observed and photographed.

Advance of Miller Glacier. In 1905, Miller Glacier showed some crevassing which we interpreted as a possible evidence of an advance, and this tentative conclusion was strengthened by the fact that the front of the glacier on the 1905 photographs from Haenke Island was apparently farther forward than in the 1899 photographs. We therefore wrote in the first draft of this report, as follows:—"It is possible that between 1899 and and 1905 this glacier had undergone an advance as a result of earthquake shaking, and that by the latter year its effects had been so healed by ablation as to be no longer noticeable. Possibly had we been on the outlook for evidence of such an advance in 1905, and had then been as familiar with the effects of an advance as we are now, we might have detected it."

We have subsequently obtained possession of a photograph taken by Ensign C. R. Miller in 1901, in which it is clearly evident that in the early summer of that year Miller Glacier was profoundly crevassed, quite in contrast to its condition in 1899 and in 1905. There can be no question, therefore, but that between 1899 and 1901 Miller Glacier was transformed by crevassing, and that between 1901 and 1905, ablation had so reduced the crevassing that it could not be certainly proved by photographs alone. It thus seems to fall in the series of advancing glaciers of the Yakutat Bay region, and about in the position in that series in which, from its small size, one would be inclined to place it—that is

[1] To it we propose to give the name Miller Glacier after C. R. Miller, who, in charge of a U. S. Fish Commission expedition in 1901, took the photograph from which we have been enabled to discover an important episode in its history.

in the Galiano series, early responding to the earthquake shaking. Like the Galiano, it must have advanced within less than two years of the earthquake of September, 1899. This evidence leads us still more strongly to the belief that, in all probability, other small glaciers of this region passed through a cycle of advance prior to 1905, and so long before it that, as in the case of the Miller and Galiano Glaciers, the evidence of the advance was quite hidden by the effect of ablation before the period of our first studies, in 1905. We look upon the discovery of the evidence of this advance of Miller Glacier, the time of which can be so definitely determined, and which falls so appropriately in its place in the series of advancing glaciers, as one of the most striking confirmations of the theory of earthquake cause for advance which has so far been discovered. It is valuable confirmation of a theory when an expected phenomenon, predicted under the theory, can be proved to have occurred.

It is further noteworthy, that the advance and transformation of Miller Glacier was less extensive than that of the neighboring Haenke Glacier. This also is expectable under the theory, for with its small size, upon which its early advance depended, a more moderate advance should result than that of the much larger and more slowly responding neighbor. The 1901 photograph does not show the exact end of the newly crevassed glacier so that we cannot now state where it terminated,

Advancing Haenke Glacier. Our evidence of an advance of Haenke Glacier is even more complete than that of Miller Glacier, for, between August, 1905, and June, 1906, it had undergone a complete transformation as was demonstrated by actual observation. The undulating, uncrevassed, moraine-covered surface of 1905 was in 1906 transformed to a perfect labyrinth of crevasses, while clear ice appeared here and there throughout the glacier from as far back in the mountains as we could see down to its front. The Haenke Glacier in 1906 was as badly crevassed as the Atrevida, but there was less clear ice showing, probably because of the slow rate of ablation here where the snow-covered mountains in the background, the ice-covered fiord in front, and Turner and Hubbard Glaciers on either side, give rise to a local climate that is much cooler in summer than in any other part of the coast of Yakutat Bay. This was proved by the presence of snow which covered the alluvial fans in this section in July, the only part of the coast of this inlet where snow remained near sea level. The crevassed surface of the glacier was, in consequence, still so black with debris that even in August we were unable to secure good photographs from a distance.

That the broken glacier had also been notably thickened is proved by a comparison of photographs. It can be seen in the earlier photographs (Pl. XLI, A) that Haenke Glacier descends over an increased slope just as it reaches the mountain front and that its lowermost portion rests on the face of this slope and spreads with decreased slope just beyond it. This condition, which closely resembles that of Turner Glacier, is interpreted, in the same way, as a result of the fact that Haenke valley is a hanging valley in its relation to Disenchantment Bay. In 1906 this increased slope was brought out with far greater clearness partly because the glacier front extended further, but largely because the amount of ice on the slope had been greatly increased. The glacier surface rose much higher, both above the lip of the ice-buried hanging valley and on its frontal slope. At the point where the glacier slope changes in passing over the lip of the hanging valley was found the greatest development of crevassing and the greatest amount of clear ice.

The most remarkable feature in the transformation of Haenke Glacier was the advance of its front completely across the fringing alluvial fan to the sea. At the same time it expanded laterally, as Turner Glacier does, with two wing-like ends, one pointing northward, the other southward. The latter wing coalesced with the north wing of Turner Glacier, thus forming a continuous ice cliff from the southern tip of Turner Glacier to the northern end of Haenke Glacier, adding nearly a mile and a half (Pl. XXXVI) to the ice cliff that had existed here ten months earlier. Although this new ice front was completely tidal there was not an active discharge of icebergs from it, possibly because of the fact that the front stood in very shallow water, resting on alluvial deposits previously laid down by the glacial stream that formed the destroyed alluvial fan. Some icebergs were being discharged, however, and the glacier front was everywhere a steep cliff; but for the most part it was so stained by the black shale moraine that clear ice appeared in it only in small patches where fragments had recently fallen (Pl. XL, B).

The forward movement of Haenke Glacier carried its front at least 4500 feet farther out than it had stood ten months before (Pl. XXXVI), and the terminus of this glacier experienced a much greater actual forward movement than any of the glaciers that advanced in 1906, but not nearly so great as the Hidden Glacier between 1906 and 1909. This greater forward thrust of the front may be due entirely to the difference in conditions surrounding the terminus. The other three advancing glaciers, the Marvine, Atrevida and Variegated (and also the Galiano) terminate in expanded piedmont ice bulbs in which the movement, thickening, and breakage resulting from the thrust dissipated itself without accomplishing a large measure of forward motion of the glacier fronts; but Haenke Glacier (and also the Hidden) had no such expanded bulb, and the thrust, even though of less degree, being concentrated on a front only slightly broader than the valley glacier itself, naturally pushed it forward. It is to be pointed out, however, that a part of the advance of Haenke Glacier may be only apparent, for it is possible that a stagnant ice mass underlay the fringing alluvial fan; but we have no proof that this was so.

Haenke Glacier from 1909 to 1913. Like the Atrevida Glacier, the Haenke since 1906 had been so healed by ablation that from a distance one would not, in 1909, have suspected that it had so recently been broken into impassable condition. The spasmodic advance cannot, therefore, have lasted much longer than the summer of 1906, if indeed it had not ceased by that time. All the crevasses were hidden and the surface was once more completely covered from side to side from the front far up into the valley, with a uniform sheet of black shale ablation moraine. The surface was, perhaps, somewhat more hummocky than in 1905, though it is impossible to prove this from the photographs, in which the condition of 1905 and 1909 seem almost the same, though in each year vastly different from the condition in the intermediate period of 1906.

The front of the glacier was apparently not so far out as in 1906, and it certainly was no longer tidal (Pl. XLI, B); but it still possessed a débris-stained cliff throughout practically its entire front. At the base of this cliff numerous débris cones and some small alluvial fans had formed, the former where moraine had slid down the ice front, the latter where streams emerged from the glacier. There were no appreciable changes between 1909, 1910, and 1913. The short time that the glacier was tidal is proof of the conclusion reached in 1906 that the ice cliff rested in shallow water. It seems safe to predict that the present cliff will soon disappear as the result of ablation, and that an extensive alluvial fan will eventually again develop and fringe the glacier front. The

glacier still expanded outside its mountain valley, as Turner Glacier does, but the southern wing seemed in 1910 and 1913 to no longer unite with the northern wing of the Turner, as it did in 1906. We could not, however, be certain of this, since the Haenke tip extends in behind the Turner and may be still united with it there.

The changes in Haenke Glacier thus agree with those observed in the other glaciers which have advanced under the influence of earthquake shaking. The advance was abrupt, spasmodic and extensive, and it was completed in a very brief period. After being observed in the summer of 1906 the glacier evidently advanced little if any more, and since then ablation has proceeded with such great rapidity and effectiveness as to heal the crevasses that the spasmodic advance produced. The notable advance of the front has not been counterbalanced by recession, and cannot be for many years.

We have given this glacier less thorough and close study than the other advancing glaciers, partly because of the difficulty of reaching it through the icebergs, but largely because prior to 1906 it had not presented phenomena of sufficient interest to warrant description and photographing from sites close enough at hand to permit more detailed comparative observations. As compared with the other glaciers that advanced in 1906 it is of interest primarily because of the very notable forward thrust of its front, which, up to the time of our observations on Hidden Glacier in 1909, was quite unprecedented in amount and rapidity. It is, therefore, rather unfortunate that we have no record of the exact position of its front before 1906.

Hubbard Glacier

General Description. Hubbard Glacier (Pl. XLII), one of the largest and grandest tidal glaciers on the North American continent, enters Yakutat Bay at the head of Disenchantment Bay. No photograph or description can do justice to the magnificence of this vast ice stream, which, having its source far back among the snow-covered mountains, flows with almost no stain of débris, as a pinnacled and crevassed flood of pure-white ice, deeply set in the mountain valleys, among lofty snow-capped peaks, and terminates in a lofty ice cliff from which icebergs are almost constantly falling. From every part of the fiord from which this glacier is visible it dominates the scene; yet, even when looking upon it from the most favorable viewpoint, and far more when looking upon a mere photograph of it, one cannot realize its size, for it is set in a mountain scene of such surpassing grandeur that even such a glacier is dwarfed by comparison. How large it is in reality may be inferred from the photograph of the glacier front on which is drawn to scale the Masonic Temple in Chicago (Pl. XLIII, A), and from the map which compares this lower part of Hubbard Glacier with the whole of the Aletsch, Mer de Glace, and Rhone Glaciers in Switzerland (Fig. 6).

Hubbard Glacier [1] has a total known length of twenty-eight miles along the north tributary, and is doubtless much longer, perhaps heading over forty miles to the north on a through-glacier divide with the Kaskawulsh Glacier north of the St. Elias Range. Besides this north ice tongue there is a broad tributary whose lower twelve and a half miles is all that has been seen by man; and there are two much narrower tributaries each at least twelve miles long, five other branches each over five miles long, and scores of smaller tributaries. The Hubbard Glacier system includes over 100 miles of known

[1] Martin, Lawrence, The Hubbard Glacier, Alaska, Popular Science Monthly, Vol. LXXVI, 1910, pp. 293–305.

valley glaciers larger than most of those in the Alps and there is, in addition, a very large unknown area.

The front of Hubbard Glacier, measured in a straight line, is 3½ to 4 miles wide (Pl. XXXVI); but since the ice cliff has a sinuous form, with projections in the center, it is in reality much longer than this, the total length being between 4½ and 5 miles. This cliff rises between 250 and 300 feet above the water and extends an unknown distance below it. One usually needs to wait but a few moments to hear from some part of the cliff the thunderlike rumble or roar which is the first announcement of an

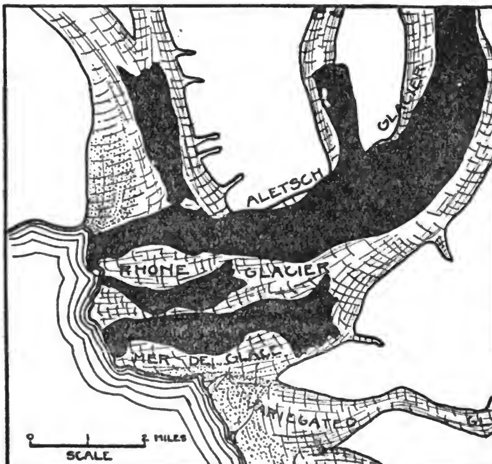


Fig 6. The Lower Portion of the Hubbard Glacier, With Three Glaciers of the Swiss Alps Superimposed Upon It.

All four glaciers are drawn upon the same scale, those from Switzerland being shown from the snowfields to the end of the ice tongue. The contrast of width and length of these glaciers in Switzerland and in Alaska is notable.

iceberg fall, followed a few moments later by the appearance of a great swell which, on reaching the shore, forms a line of white breakers even at a distance of several miles from the ice cliff. By watching the ice cliff, one may see the huge masses fall from the ice front and a fountain of water dash perhaps even to the top of the glacier, and then, in a few seconds, hear the report which the rending of the glacier sends out. One is fascinated by the performance; sometimes it is only a small piece that falls and then a sharp single report, like the crack of a pistol, goes through the air; again a part of the front crumbles and the down-sliding ice, broken into small pieces, seems from a distance like a fountain of water while the report is only a low rumble; at other times huge masses break away, forming large icebergs, and the noise then produced is like the heavy

rumble of distant thunder; and at rare intervals one may see a huge mass of blue or black ice thrust itself up from below the fiord, some distance from the glacier front, as a part of the submerged ice foot is broken off, and then no report is heard, but the wave that follows is far greater than usual.

The water waves which follow the discharge of icebergs from the front of Hubbard Glacier are of great magnitude. In September, 1913, for example, the steamship *Princess Maquinna* was aground on a reef just west of Osier Island with the members of the International Geological Congress who visited Yakutat Bay under the guidance of the junior author. The ship lay over a mile from the glacier, and yet the water wave following the discharge of icebergs from Hubbard Glacier caused the steamer to roll until she took in water on the main deck, as first one side and then the other was tipped far down by the iceberg waves.

There are periods when for an hour or two there is very little discharge, and then periods, fully as long, when scarcely a moment elapses without the sound of ice-falls from some part of the glacier front. It is possible that these differences are in some way related to the state of the tide, but we have not determined whether the periods of quiet and activity form part of a regular cycle or are merely irregular intervals due to accumulation of strain and relief from strain or to expansion and contraction under sunlight. While there are periods of relative quiet, they are not periods of absolute repose, and they occupy far less time than the periods of activity. Day and night the ice falls and the reports that pass out through the air are so frequent that it is fair to speak of the glacier as almost ceaselessly active. The noise disturbs one's sleep at first and sometimes, when an unusually heavy fall occurs, wakens one even after he has grown accustomed to the ordinary rumble. A sense of nervous relief is felt when camp is removed to a part of the fiord to which the iceberg roar and the breakers on the coast do not reach.

This almost ceaseless activity of iceberg discharge testifies conclusively to the activity of the glacier behind. To supply so much falling ice (Pls. XLII, XLIV) there must be rapid movement up to the front. The broad stream of floating ice that stretches throughout Disenchantment Bay, and down Yakutat Bay to a distance fully 15 miles from the glacier front (Pl. L), offers further testimony in the same direction. And back of the ice cliff, where the surface from side to side is shattered by an impassable complex of yawning crevasses, sharp ice pinnacles, and seracs, one sees almost equally impressive evidence of rapid movement. The surface of Hubbard Glacier offers a strong contrast to that of the slowly-moving and stagnant ice masses, but resembles their condition during their brief periods of spasmodic advance. How fast the ice is moving is unknown, for no observations of rate have yet been made, but all facts indicate that, as glaciers go, it is a very rapidly moving ice stream.

Concerning the reservoirs which supply this immense ice stream, and most of the tributaries which unite to form it, we have no further knowledge than that supplied by the Boundary Commission map.[1] According to this map the sources of the Hubbard Glacier are far back among the lofty mountains of the St. Elias range, and probably in an ice-flooded area similar to that described by Russell as visible from the upper slopes of Mount St. Elias. No one has yet explored this region and the difficulties in the way of such exploration are great, though probably not impossible. We have been able to look far up the glacier, and as far as we could see, many miles from the coast, there is a

[1] Atlas of Award, Alaskan Boundary Tribunal, Sheet 23.

broad valley filled with crevassed ice, and bordered by lofty, snow-clad mountain peaks, one of which, Mount Hubbard, rises to an elevation of 16,400 feet at a distance of 30 miles from the Hubbard Glacier front. In the absence of a more thorough survey we cannot be certain of the length of Hubbard Glacier, but from the evidence that we have the distance from its front to the head of some of its tributaries cannot be less than 30 miles and may be even as much as 50 miles. Doubtless it is made by the union of many tributaries of various sizes from Mount Hubbard, Mount Vancouver, and the mountains behind. It is probable that some of these upper portions of the Hubbard are through glaciers, feeding not only the Hubbard but also glaciers descending to the inland side of the mountain range, and possibly even directly connected with some of the feeders of Malaspina Glacier on the west, as they are known to be with Nunatak Glacier on the east.

The outer portion of Hubbard Glacier, with its ice cliff, is made by the union of two large arms, one from the direction of Mount Vancouver to the northwest, the other and larger from the Mount Hubbard region to the north. The northwest arm is a mile and a half or two miles wide, the north arm from two to three miles wide where it emerges from its mountain valley. Within the area visible from points on the fiord several relatively small tributaries, extend to each of these arms, descending the mountain valleys with steep grades and in some cases as cascading glaciers, or else with cascading ends just above their junction with the main glacier. Although small in comparison with the two main arms of Hubbard Glacier several of these tributaries are comparable in length and width with the ordinary valley glacier of the Alps. Each of the main arms of the glacier has a moderate grade within its mountain valley, the surface slope just inside the mountains being estimated to be about 5°; but where they emerge from the mountains, and just above where they coalesce, the grade abruptly increases to twice that amount. As in the case of the Turner Glacier this steepened slope is interpreted as the result of the fact that the mountain valleys which the two arms occupy are hanging in relation to the main fiord valley. In the same way is the cascading condition of the lower ends of the tributaries to the two arms of the Hubbard interpreted as a result of the fact that these tributary valleys are hanging above the valleys in which the two main glacier arms lie.

Below the steepened slopes the north and northwest arms unite, forming a broad plateau with an undulating but nearly horizontal surface, broken by a labyrinth of crevasses and bristling with ice pinnacles, and faced by the ice cliff already described. Whether any part of this terminus is afloat, or whether it all rests on the rock floor of its valley cannot be stated. At the base of the steepened slope the ice surface is only a few hundred feet above the fiord level, warranting the prediction that, if Hubbard Glacier should retreat, Disenchantment Bay would be extended several miles, and probably up to the base of the steepened slope.

Both the two arms and the low-lying ice plateau which their union makes are remarkably free from morainic débris (Pl. XLVI). This is in the main doubtless due to the fact that their rapid motion prevents the work of ablation from proceeding far enough to concentrate débris on the surface before reaching the front, where both ice and débris are floated away. A contributory cause for the absence of débris is the broadness of the valleys, by reason of which the avalanches in the lower valley portion cannot spread out toward the center of the glacier, while those that fall higher up are not revealed by

ablation. Another cause for the clean surface is the abundance of crevasses into which rock fragments would naturally fall. While there is little moraine on the surface, the abundance of black, dirt-laden icebergs that float away from the front of Hubbard Glacier prove that there is much débris in the lower layers; and on the border of each arm there are well-defined lateral moraines, while the northwest arm has also a medial moraine. That the north arm has no medial moraines can only be explained by its rapid motion and the consequent ineffectiveness of ablation, for it is inconceivable that so great a glacier, heading so far back among lofty mountains, can fail to have tributaries of sufficient size to contribute notable supplies of medial morainic débris. It is to be noted, however, that close by each lateral moraine, and only a short distance out from it, is a moraine ribbon that probably represents a medial moraine of some tributary, pushed to one side by the dominance of the other arm of the main glacier.

Of the two large arms visible from the sea, the northern is clearly not only the largest but far the most active, and, in fact, dominant in the ice plateau that is made by the coalescing of the two arms. There are several reasons for reaching this conclusion. The first of these is the larger size of the north arm. Even more noteworthy than this is the position of the moraines in the outer ice plateau. The two arms which form the plateau approach each other nearly at right angles, but the dominance of the north arm is made apparent by the fact that its western lateral moraine proceeds in fairly direct course to the fiord, whereas the medial and lateral moraines of the northwest arm bend sharply and extend to the fiord parallel and close to the lateral moraine of the north arm. The northwest arm is far too weak to deflect the lateral moraine of the north arm, but the north arm is strong enough to bend the moraines of the northwest arm at a right angle to the course which they were following up to the time that they came under its influence (Pl. XLV).

The dominance of the north arm in the terminal ice plateau is also shown by the proportion of ice which it contributes to the terminus. After descending its steepened slope, this arm fans out slightly, and at the glacier front supplies the ice for fully two-thirds of the ice cliff. This conclusion is based upon the distance between the eastern lateral moraine and the moraine which comes down from the northwest arm. We estimate that the width of the ice front dominated by the north arm is fully one-half greater than the width of the north arm where it emerges from its mountain valley over the steepened slope.

In addition to these evidences of the dominance of the north arm, the northwest arm furnishes proof of its own weakness. It has much more moraine on its surface than the north arm, including a medial moraine, and this is interpreted as the result of its slower motion and the resulting greater effectiveness of ablation. Furthermore, the northwest, arm seems to have a buried irregularity of the valley floor, revealed by a slight bulge where the glacier comes out of its hanging valley. The débris-covered ice below the bulge, for which there is no visible source, suggests a ledge near the surface, for this is the only dirty portion of the glacier away from the moraines and may be seen in earlier photographs by Russell and Gilbert. Its character is seen well from near Turner Glacier. Since this area occurs out in the glacier, at some distance from the margin, it is interpreted as evidence that the ice in the northwest arm of Hubbard Glacier is not as thick where it passes over its steepened slope as the north arm is on its hanging valley lip.

The Southeastern Margin. Partly because of the interesting features which it presents,

and partly because of its relation to Variegated Glacier, the eastern margin of the Hubbard Glacier was studied in some detail in 1905; and in view of the fact that notable changes had taken place when we studied it again in 1909, its characteristics will be stated here in some detail. Within the area of observation this margin may be divided into three portions, an upper, middle and lower portion.

In the upper portion, which lies within and just outside of the mountain valley, there is a border of lateral moraine of considerable breadth, which near the very margin is so thick that most crevasses are filled, making it possible for one to travel over the surface. But a short distance out on the ice the surface becomes badly crevassed and the moraine much thinner. Between the glacier and the mountain there is a marginal valley in which a small stream flows, and this is continued westward by a low, undulating depression in which lie several marginal lakes (Pl. XLVII, B.) This depression is an interlobate area between the Hubbard and Variegated Glaciers, and its outer portion, where the two glaciers coalesce, is underlaid by ice.

Below the depression, for a distance of about two miles, the Hubbard and Variegated Glaciers are united and one cannot state exactly where the line between the two should be drawn (Pl. XLII). Nevertheless it is perfectly plain that all but a very small area along the exact boundary is to be assigned to one or the other of the glaciers. This is made evident by the character and distribution of the moraines. Hubbard and Variegated Glaciers approach each other at very nearly right angles, but Variegated Glacier spreads out in a broad, stagnant bulb, covered with moraine from side to side, while Hubbard Glacier pursues its course to the sea as a vigorous, actively-moving glacier, uncrevassed and moraine-covered only on the very margin. The marginal morainic ribbons on the Hubbard bend as if the Variegated dominated; but, in view of the stagnant state of the Variegated, this cannot be the case and we must assume that the Hubbard is set in a valley of sufficient depth to deflect it westward and prevent it from invading the area occupied by the stagnant Variegated, toward which it is directed when it comes from its mountain valley. On the two sides of the area of coalescence the moraines are wholly different. On the Hubbard side there are bands parallel to the direction of ice motion toward the fiord and in these morainic bands fragments of red gneiss are sufficiently predominant to give the moraine color; but on the Variegated side the ribbon appearance is absent and in its place there is a confused waste of ablation moraine blackish in color because of the great abundance of black hornblende gneiss fragments, and with a roughly-crescentic banding apparent when viewed from a distance.

There is a noteworthy difference in glacier condition on the two sides of the area of junction of the Variegated and Hubbard Glaciers. Where the two unite the glacier surface is so deeply covered with moraine that the ice is in large part hidden, though it appears here and there in crevasses and in cliffs. The emergence of springs, the abundant evidence of slumping of the moraine, and the general absence of vegetation all testify to the presence of ice beneath those areas in which no ice is visible. Usually we could also prove the existence of ice in such places by thrusting our ice axes down to it through the morainic veneer (Pl. XLVII, A). The irregular protection which the moraine cover offers to the underlying ice here, as elsewhere, gives rise to the development of hillocks and ridges, with intervening kettles and valleys. This condition of the glacier extends not only over the junction of the two glaciers, but also southward over the stagnant outer bulb of the Variegated Glacier.

Toward the north, on the other hand, in the region dominated by the northern arm of the Hubbard, the condition is wholly different. Crevasses appear, the moraine no longer forms a continuous cover but exists in ribbons with intervening areas of relatively clean ice, and in a very few yards the glacier surface becomes quite impassable. Beyond lies the great sea of clean, white, crevassed and pinnacled ice, contrasting strikingly with the undulating black moraine of the Variegated Glacier. There is much variety in detail in this zone of actively-moving marginal ice of the Hubbard. For example, at one point the broken ice forms a pronounced dirt-covered ridge rising above the neighboring clear ice. Here the layers are steeply-inclined, and it is evident that lower débris-charged layers have been upturned and that the débris has so protected the broken ice from ablation that it stands up as a prominent, crevassed ridge. Here and there are small ponds on the ice surface, and everywhere there is great variety in the crevassing. One very notable feature in 1905 was the presence of a saucer-shaped depression, fully 250 yards in diameter and 100 feet deep, in the actively-moving, crevassed ice, giving a local down-stream ascent to a part of the glacier surface just beyond the steep descent from the mountain valley. There are other undulations of the marginal surface which are interpreted as indications that the glacier is flowing over an irregular bottom.

From this description the relations of the Hubbard and Variegated Glaciers are clear. They coalesce for some distance, but each maintains its individuality in all but a very narrow strip, and one is vigorously active, the other quite stagnant. It is a question whether the united ice mass from the northwestern margin of the Hubbard to the southeastern margin of the Variegated ice bulb should be considered as one piedmont glacier. In a sense this is a warranted interpretation of the phenomena, and both Russell and Gilbert adopted this interpretation, and extended the name Hubbard Glacier to include not only the actively-moving ice but also the stagnant Variegated bulb. We are inclined, however, to confine the name Hubbard to the part dominated by the vigorous north and northwest arms, and apply the name Variegated to the stagnant piedmont bulb which it has formed, because the two glaciers are so different in character, and each portion is completely formed and dominated from a different source. At the same time, we recognize the fact that this entire low-lying ice mass, outside the individual valleys from which the supplying glaciers issue, is a continuous ice plateau, analogous to a piedmont glacier, though itself within a mountain valley. In this sense it might be warranted to give a name to the ice plateau, as in the case of Malaspina Glacier, and to give separate names to the three contributing glaciers—the northwest arm, the north arm, and the Variegated. A fourth glacier, the Orange, which barely coalesces with the Variegated, is to be classed with this system. Certainly at no very distant date, as is proved by the high-lying moraine terrace and the marginal drainage channels, these four glaciers contributed to the formation of a great trunk glacier which flowed out into Russell Fiord and Disenchantment Bay; but now recession has proceeded so far that the two eastern arms are so separated from the Hubbard that we believe it best in accordance with the facts to treat them as separate glaciers down to their terminus.

The third portion of the southeastern margin of Hubbard Glacier is the outermost part, beyond the point where the glacier turns so far to the west that it is again separated completely from the Variegated bulb. In this portion the two glaciers are separated by a V-shaped valley in whose bottom a small, branching glacial stream flows, depositing outwash gravels (Pl. XLIX). From the sea to the head of the valley the distance is about

a half mile, and at its head the undulating, moraine-covered ice appears, with a depression between the two glaciers for several hundred yards further. On the south side of the valley rises the outer portion of the Variegated ice bulb, here low, undulating, and deeply covered with moraine. On the northern side (in 1905 and 1906) the Hubbard Glacier rose more steeply, but still with low enough slope so that one could easily ascend it. Moraine completely covered this wall of the glacier and no ice was visible in it; but just north of it, on the Hubbard Glacier surface, the moraine became thinner and more crevassed, and in a short distance one came to the normal broken surface of the Hubbard. This moraine-covered margin extended to the sea and there rose as a débris-stained cliff, becoming rapidly higher and clearer of débris toward the west until the normal white ice cliff of the Hubbard front was reached. The amount of crevassing also increased away from the margin, and with it the amount of ice discharged. For a quarter of a mile from the glacier margin the débris-stained ice cliff was so inactive in 1905 and 1906 that iceberg discharge was practically absent. Only one or two small fragments were seen to fall, and the area of débris-stained cliff face was far greater than the area whitened by recent falls. Moreover, there was pronounced undercutting of the ice cliff by the sea water which proved clearly that iceberg discharge was not rapid. These facts are of importance by contrast with the condition in 1909 when an advance had begun along this part of the southeastern margin of the Hubbard Glacier.

Hubbard Glacier Before Russell's Studies. In 1792 and 1794 Malaspina and Vancouver visited Yakutat Bay but tell us practically nothing specific concerning Hubbard Glacier. Malaspina's map, however, and the description by Malaspina and Vancouver were interpreted by Russell,[1] Gilbert,[2] and Davidson [3] as proving that Hubbard Glacier, joined by Turner, then extended southward to a point south of Haenke Island and five or six miles south of where it ends now. The authors of this book have opposed this interpretation,[4] because of evidence from existing vegetation and shorelines and absence of lacustrine deposits and of abandoned shorelines in a large part of Russell Fiord, and we place a different interpretation upon the conditions encountered by Malaspina in the light of existing floating ice phenomena, although we recognize the presence, rather recently, of a smaller lake in lower Russell Fiord.

In Tebenkof's Atlas of 1848 the chart made, as Davidson has shown, by the Russian Booligin in 1807 and Lieut. Khromtchenko in 1823 has a crescentic ice front a short distance north of Haenke Island and a lake is shown in Russell Fiord. It is five or six miles long and there is no independent lake now. It is not dissimilar in width to upper Russell Fiord. It drains southward to the Pacific by the Situk river which is a little too long on the map and which now rises near the present head of Russell Fiord. The former glacial lake in the head of Russell Fiord, whose shorelines we have studied, was of almost exactly this length and its outlet was probably the Situk.

The front of Hubbard Glacier may have been north of Haenke Island in 1823 and the expanded Nunatak-Hidden Glacier reached the south part of Russell Fiord in that year

[1] Russell, I. C., Nat. Geog. Mag., Vol. III, 1891, pp. 67, 97–98; 13th Ann. Rept., U. S. Geol. Survey, 1894, p. 84.

[2] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, p. 70.

[3] Davidson, George, Trans. and Proc. Geog. Soc., Pacific, 1904, p. 51.

[4] Tarr, R. S. and Martin, Lawrence, The Position of Hubbard Glacier Front in 1792 and 1794, Bull. Amer. Geog. Soc., Vol. XXXIX, 1907, pp. 129–136.

and no conditions of vegetation known to us are opposed to this interpretation. The rapid retreat of these glaciers to their present positions is not impossible, judging by the rapid retreat of Nunatak Glacier from 1895 to 1909 and the recession of Muir and Grand Pacific Glaciers in Glacier Bay a mile and a half a year[1] from 1794 to 1912. The absence of shorelines in northern Russell Fiord might be accounted for by the rapidity of retreat and quick shifting of lake levels, or by retreat southward instead of northward as suggested in a later paragraph.

In 1888 Topham reports[2] that "at the northeast end of Yakutat Bay the sea-water flows inland through a narrow passage frequently blocked by ice into a lake known as Disenchantment Bay. This bay is said to be 30 miles long and to be surrounded by high mountains and great glaciers." In 1890 Russell went far enough north of Haenke Island in the *Corwin* to see beyond the bend into Russell Fiord which he explored to the head in 1891, finding Hubbard Glacier not far from its present position.[3]

The authors are still of the opinion that Russell, Gilbert, and Davidson misinterpreted Malaspina's map and description. We believe, however, that Hubbard Glacier may have extended some distance farther south in 1792 than now, ending nevertheless north of Haenke Island then and in 1794, 1807, and 1823. We are inclined to believe that the Russell Fiord Lake existed as recently as 1823, as the map shows specifically, although this implies that an east branch of Hubbard-Variegated Glacier, joined by the Nunatak and Hidden Glaciers, then extended over twenty miles southeast and south to within six miles of the present head of Russell Fiord. The evidence is conclusive that this condition actually existed in a very recent period, and it may have been as late as 1823.

The Disenchantment Bay bifurcation of such an expanded ice tongue would be farther back in 1823 than the Russell Fiord branch because the latter ended temporarily in fresh water rather than salt, but soon melted back and was dismembered because salt water from the Disenchantment Bay side got at it. A Yakutat native told us in 1905 that his father-in-law, then still living, remembered when Nunatak Glacier extended to a point near Marble Point, about five miles southeast of Hubbard Glacier in Russell Fiord. This all accords well with the idea of a great advance of all these glaciers in the late eighteenth or early nineteenth century, the Hubbard and Nunatak Glaciers having now melted further into the moutains than the adjacent Yakutat and Malaspina, because outwash deposits protect large parts of the last two from salt water and rapid melting.

Changes Between 1891 and 1905. From several points of view the Hubbard Glacier was photographed by Russell in 1891 and by Gilbert in 1899, and both of these observers have described the glacier at the time of their observations, while Gilbert has made a study of the conditions in 1899 as compared with those in 1891. Photographs were also made by the Canadian Boundary surveyors in 1895, by Bryant in 1897, and by the U. S. Fish Commission in 1901. In 1905, and again in 1906, we occupied the exact stations from which some of Russell's and Gilbert's photographs were taken, and the comparison of conditions was described in our earlier reports.[4] Recent retreat

[1] Martin, Lawrence, Glaciers and International Boundaries, Scientific American Supplement, Vol. LXXVI, 1913, pp. 129, 136–138.

[2] Topham, H. W., Proc. Roy. Geog. Soc., Vol. XI, 1889, p. 425.

[3] Russell, I. C., Nat. Geog. Mag., Vol. III, 1891, pp. 99–100 and Pl. 9; 13th Ann. Rept., U. S. Geol. Survey, 1894, pp. 85, 89.

[4] Tarr, R. S. and Martin, Lawrence, Glaciers and Glaciation of Yakutat Bay, Alaska, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 154–155; Tarr, R. S., Professional Paper 64, U. S. Geol. Survey, 1909, p. 45.

southeast of the glacier is proved by morainic margins so recently abandoned as to be free from vegetation.

Gilbert had previously shown that in 1899 [1] Hubbard Glacier had retreated one or two hundred feet since Russell made his studies in 1890 and 1891, as viewed from Osier Island and slightly since 1895 as viewed from Haenke Island. The retreat was even greater between 1899 and 1901, for Gilbert states that the 1901 pictures by the Fish Commission party show that "at a point where a prominent moraine makes the comparison somewhat definite, the ice cliff appears to have stood 700 to 1000 feet farther back than in 1899. The cliff was also shortened at each end by the enlargement of the marginal belts of stagnant ice."

Between 1899 and 1905 the northwest part of Hubbard Glacier advanced, as shown by comparison of our photographs with those from Gilbert's photographic site above Osier Island. The west margin was farther out in 1905 than when photographed from Osier Island by Russell in 1891. It was several hundred yards farther out than when photographed from Haenke Island by Russell in 1891 and by Gilbert in 1899. The bend in the medial moraine of the northwest arm, as seen from Haenke Island and from the mountain side above Osier Island, was a quarter of a mile farther southeast in 1905 then in 1891, and several hundred yards farther in 1905 than in 1899.

The exact site of Gilbert's 1899 pictures of Hubbard Glacier from the coast between Variegated and Hubbard Glaciers could not be found in 1905 because altered by wave work; but from approximately the same site, that is with the same points in line on the mountains in the background, it was perfectly clear that the southeast side of Hubbard Glacier had retreated between 1899 and 1905, probably several hundred yards.

This advance of the northwest part of Hubbard Glacier and retreat of the southeast side between 1899 and 1905 we interpret as an indication of an advance of the weaker northwest arm under the impulse of the earthquake shaking, the strong north arm retreating relatively because of ice lost during the severe shaking of the 1899 earthquakes, as proved by the direct testimony of prospectors who were encamped beside it in 1899 [2] and by the Fish Commission pictures from which Gilbert proved a retreat of 700 to 1000 feet between 1899 and 1901, as already stated.

Condition in 1906. In this year we thought that there was a more active discharge of icebergs from Hubbard Glacier front than in the previous season, and that there was a greater quantity of ice floating in Disenchantment Bay. But such evidence is of little value, being mere impression. Comparison with the photographs of 1905, however, gives evidence of a slight advance in the northwestern portion of the ice front, but nothing to compare with the advance of other glaciers in the region, including the smaller Variegated and Haenke Glaciers nearby. We made a visit to the southeastern margin of the Hubbard and found the condition there unchanged, so that the description given in previous pages applies as fully in 1906 as to the preceding year.

Advance in 1909. In the summer of 1909 evidence was observed of a very distinct forward movement (Pl. XLVIII), though it was apparently only the beginning. This

[1] Gilbert, G. K., Glaciers and Glaciation, Harriman Alaska Expedition, Vol. 3, 1904, pp. 63–66.

[2] Tarr, R. S. and Martin, Lawrence, Recent Changes of Level in Alaska, Bull. Geol. Soc. Amer., Vol. XVII, 1906, p. 31; Martin, Lawrence, The Alaskan Earthquakes of 1899, Bull. Geol. Soc. Amer., Vol. 21, 1910, p. 359; Tarr, R. S. and Martin, Lawrence, The Earthquakes at Yakutat Bay, Alaska, in September, 1899, Professional Paper No. 69, U. S. Geol. Survey, 1912, pp. 15–17.

evidence was so clear that it is brought out distinctly in the photographs taken from the earlier sites. From Gilbert's site, for instance, at an elevation of 1000 feet on the mountain slope above Osier Island, it was clearly evident that the northwestern arm had continued to advance since 1906, the extreme western margin being several hundred yards farther out and the ice cliff in that part of the glacier being slightly extended. One of the evidences of advance of the northwest arm discovered in 1905 was changed, namely, the position of the bend of the medial moraine of that arm. This bend was further back, and where deflected westward, the curve was blunter than in 1905. This might be due either to increase in weakness of the northwest arm or to increase in strength of movement of the north arm. That it is not the former is proved by the continued advance of the western margin of the Hubbard. This leads to the acceptance of the latter interpretation, and direct proof of its correctness is supplied by the condition of the southeastern part of the glacier. The extreme outermost part of the ice front, which is supplied from the north arm, projected 200 or 300 feet farther out than in 1906 and at least a quarter of a mile farther than in 1891. Along the southeastern margin there was much less débris than in 1905 and 1906 and the eastern portion of the ice cliff had a far larger proportion of clear ice. A further noteworthy fact was the position of a morainic ribbon that lay on the glacier just outside of the southeastern lateral moraine. In 1906 it was much farther out in the glacier than in 1909 as if in the meantime it had been pushed westward, in toward the valley wall by the advance of the glacier.

All these facts convince us that while the northwest arm had advanced somewhat, the north arm had also begun an advance in the interval between 1906 and 1909. There was quite certainly a much more active discharge of ice from the glacier front than in 1905 and 1906. Sometimes there was an almost steady roar for a half hour. There was very heavy ice off the glacier front, far more than in former years, though it is of course true that this condition varies greatly from time to time. The natives report such heavy ice in 1909 as to interfere with their sealing. Neither in 1905 nor 1906 was there nearly so much stranded ice on the beach at our camp site on the west shore of the bay as in 1909. It seemed certain that the forward thrust of the ice front, which was unquestionably in progress, was accompanied by a more rapid discharge of ice into the fiord.

Clear and convincing proof of recent advance was found at the southeastern margin of the Hubbard Glacier. There was certainly a considerable advance of the front as viewed from the photographic sites selected by Gilbert in 1899 and re-occupied (approximately) by us in 1905 and 1906; but since these sites are on the beach we cannot be certain that we re-occupied them exactly, and accordingly, in 1909, we established a new photographic station (F. Map 3) on a neighboring morainic mound, marking its site by a cairn.

The clearest evidence of advance of the southeastern margin, however, is furnished by the change in the condition of the glacier itself. The ice cliff, which extends here from the land into the sea was much more broken than formerly, it had a much smaller proportion of débris-stained ice, and there were frequent falls of icebergs from its face. In former years only small blocks fell from this front, and then only rarely, but during the period of our observation in 1909 there were many ice falls, some of them of large size. This indicated that the stagnant margin was advancing and there was abundant further proof of this conclusion in the change in condition of the margin where it rested on the land nearby. In 1906 and previous years this margin was a uniformly débris-covered

slope up which one could easily climb. In 1909 the margin was in the main an ice cliff, too steep for easy climbing, rent by a number of recently opened cracks, and broken by thrust faults. The extreme recency of this breaking was made evident by the angularity of the edges of these crevasses, which had not been exposed long enough for ablation to greatly modify the original angularity of breakage.

A veneer of from 1 to 3 feet of moraine rested on the upper portion of this newly-exposed ice cliff, and in it occasional young willows were growing. It had the appearance of an ice cliff recently steepened, broken, and protruded through the moraine cover, the change being easily seen by comparison of photographs made in 1905 and 1909. As the ice cliff melted, the moraine from above was sliding down, and with it the willows that were thus undermined. These rock fragments with a few bushes, together with the débris that had fallen away from the steep ice face, had formed a small talus at the base of the ice cliff. For fully half a mile bowlders, clay, and other morainic débris were cascading down the steep ice face.

From the new ice cliff many small, muddy streams were issuing and uniting, forming a muddy glacial stream of considerable proportions in the V-shaped valley between Hubbard and Variegated Glaciers, where previously there was only a small stream, from the head of the valley, with trickling, clear-water tributaries from the Hubbard margin. This augmented drainage united to form a muddy pond back of the bowlder beach, through which it escaped in one fairly large channel. Where this pond stood, and between it and the glacier margin, annual and perennial plants and willow bushes throve in 1905 and 1906, but in 1909 many of these were submerged either in the icy water or in the mud that it was depositing. On one willow, around which we dug in the freshly-deposited mud, we found bark over a foot below the surface, from which new roots were starting. Practically all the willows partly submerged in mud were still growing, which is taken as evidence that the deposits which were overwhelming them had been there but a short time.

Continued Advance in 1910. Between July, 1909, and June, 1910, the junior author found that forward movement of the northwest side of Hubbard Glacier (Pl. XLVIII) exceeded that in the southeast, where some points actually receded, although at least two points on the south coast advanced during this period and the crevassing near the margin increased decidedly. The specific measurements quoted below are based upon a resurvey of the ice front originally plotted upon the plane table map in 1909 (Map 3). The photographs taken in 1910 also show the changes when compared with 1909 photographs from the same sites. From Osier Island (Photo. Station C, Map 3) it was clear that there had been an advance of the ice point just west of the center. From Gilbert Point (Photo. Station D, Map 3) a very slight retreat of the west edge of the glacier was shown, the ice point next it in the northwest part of the glacier had advanced, apparently between six and seven hundred feet, the ice point nearest midglacier was either stationary or showed very slight retreat, an adjacent cliff (the one projecting farthest in 1909) retreated nearly 1000 feet, a point of black ice nearer the southeast side of the glacier was shortened almost 500 feet and large areas of black ice on the east side were diminished or had completely disappeared. From Haenke Island (Photo. Station A, Map 3) it was clear in 1910 that the northwest portion and the east central cliff had advanced slightly since 1906. This photograph station was not visited in 1909.

Plate XXXIII

Landslide into Marginal Stream of Galiano Glacier in 1909

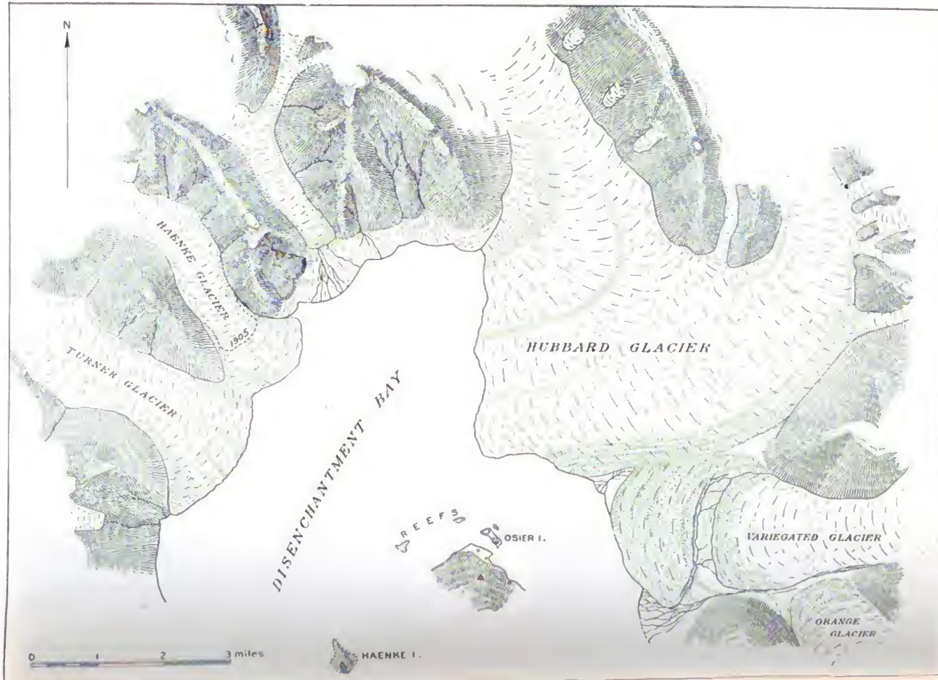
A.

B. Vegetation on the Alluvial Fan and Moraine of Galiano Glacier
In 1905 (upper picture) and in 1909 (lower picture).

A. Trees Killed by Glacial Torrent of Galiano Glacier in 1905

B. Black Glacier in 1890
Photograph by I. C. Russell from Station G (Map 2).

Plate XXXVI


N
HAENKE GLACIER
1905
TURNER GLACIER
DISENCHANTMENT BAY
HUBBARD GLACIER
REEFS
OSIER I.
HAENKE I.
VARIEGATED GLACIER
ORANGE GLACIER
0
1
2
3 miles

Turner Glacier from Crest of Haenke Island

Glaciated surface in foreground. Haenke Glacier, terminating back from the fiord, with snow-covered alluvial fan in front, just to right of Turner Glacier. Photograph by A. J. Brabazon, Canadian Boundary Commission, 1895, from Station A (Map 3).

Plate XXXVIII

Turner Glacier Cliff in 1906
Compared in height with a lofty office building in New York City.

A.

B. Southern Margin of Turner Glacier in 1905 (upper view) and in 1909 (lower view)
From Station B (Map 3). Note the increase in morainic covering.

Plate XL

A. Turner Glacier in 1910
From Station C (Map 3) on Osier Island.

B. Haenke Glacier in 1906
From same site as upper view. It was then crevassed and advancing (see Plate XLI, B).

A. HAENKE GLACIER IN 1899

From Station A (Map 3) on crest of Haenke Island (see Plate XXXVII). Photograph by G. K. Gilbert.

B. HAENKE GLACIER IN 1909

From Station C (Map 3) on Osier Island (see Plate XL, B).

PLATE XLII

HUBBARD GLACIER IN 1909
From Station D (Map 3) on Gilbert Point south of Osier Island and 1000 feet above sea level (see Plate XLVI).

PLATE XLIII

A. THE HUBBARD GLACIER CLIFF
Compared in height with the Masonic Temple in Chicago.

B. ICEBERG IN YAKUTAT BAY IN 1905

A. Iceberg in Yakutat Bay

B. Icebergs in Yakutat Bay

HUBBARD GLACIER. JUNCTION OF ITS TWO MAIN BRANCHES
Photograph by A. J. Brabazon in 1895. Mt. Vancouver on left.

PLATE XLVI

HUBBARD GLACIER. FRONT AND NORTHWEST ARM
Photographed in 1910 from Station D (Map 3), on Gilbert Point, south of Osier Island (see Plate XLII).

A. Stagnant Moraine-Covered Eastern Margin of Hubbard Glacier
Slumping so rapidly that no vegetation grows. Photograph taken July 5, 1905.

B. Eastern Margin of Hubbard Glacier (on right)
Near emergence from mountain valley. Showing swinging of lateral moraines westward away from the Variegated Glacier bulb. Photograph taken July 5, 1905.

PLATE XLVIII

THE HUBBARD GLACIER IN 1909 FROM OSIER ISLAND

From Station C (Map 3). The white lines show positions of the ice front as photographed from the same station between 1891 and 1910.

From the cairn built in 1909 between Hubbard and Variegated Glaciers (Photo. Station F, Map 3), it was plain in 1910 that one point near the center of the glacier had advanced since 1909 perhaps several hundred feet, that the dirty projecting cliff had advanced slightly, and that crevassing in the non-tidal southeast edge of the glacier had increased somewhat. The same was observed from near the site of Gilbert's beach station, close to the ice edge (Photo. Station E, Map 3). The area of black crevassed ice bared by June, 1910, and of cracking had not increased so much near the head of the stream between Hubbard and Variegated Glaciers as close to the tidal portion, though even here the additional crevassing was not remarkable. The surface of the margin of the glacier still showed many minor hummocks of ablation moraine that were present in 1909. The dirty tidal cliff was a little less débris-stained than eleven months before. All in all, therefore, although there had been noticeable change, the transformation was not remarkable. One could say distinctly that advance had continued since 1909, but it was no such advance as took place in Atrevida, Variegated, Haenke, and Marvine lobe of Malaspina Glacier between 1905 and 1906. Indeed, with the specific measurements of retreat shown in some points of the ice front (Map 3), it is thought possible that the advance initiated in the southeast portion of the glacier between 1906 and 1909 was nearly over in 1910.

The advance of the northwest portion of the glacier which began earlier seemed to be continuing more vigorously. Iceberg discharge from the front of Hubbard Glacier was probably less active in June, 1910, than in the previous July. More floating ice was present, but that was thought to be due chiefly to another cause.

Condition in 1913. Because of the delay incidental to the detention of the International Geological Congress ship on the newly-discovered reef near Osier Island in September, 1913, the junior author was unable to reoccupy any of the photographic sites established in earlier years. It was clear, however, that the slight advance of Hubbard Glacier in 1909–10 was practically over before 1913. No part of the ice front had advanced significantly and the western and southeastern margins seemed to have retreated slightly. Iceberg discharge was much less active than in 1910, a fortunate circumstance to which may be ascribed the fact that no more damage was done to the vessel during her half day's stay upon the reef in front of Hubbard Glacier.

Significance of the Advance. The complex history presented by the Hubbard Glacier corresponds, in a way, with that of the Turner, and the interpretation placed upon it is the same. It does not follow, of course, that the rather deliberate advance of the northwestern arm from 1901 to 1910 following the retreat of 100 to 200 feet between 1890 and 1899 and of 700 to 1000 feet between 1899 and 1901, is the only response that this part of the glacier will show to the effects of earthquake shaking. It may be that the advance was merely the result of climatic influences, or of an impulse due to earthquake shaking imparted by one or a small number of tributaries, and that a still greater forward movement is yet in store. It would be interesting and quite unexpected if so large a glacier gave such a weak response to the cause which pushed the front of the far smaller neighboring Haenke Glacier forward nearly a mile, or if the north arm, which is so active, should fail to respond strongly and advance vigorously later, as it did weakly in 1909–10. It is by no means an impossibility that, under the influence of such a thrust, Hubbard Glacier front might be pushed forward two miles across the fiord to the mountain side behind Osier Island, making Russell Fiord into a lake once more.

TABULAR STATEMENT OF OSCILLATIONS OF HUBBARD GLACIER

<table>
<tr><th>Year</th><th colspan="2">Nature of Change</th><th>Based on observations by</th></tr>
<tr><td>1890–1891</td><td></td><td rowspan="4">Slight advance of west part</td><td>Russell</td></tr>
<tr><td>1891–1895</td><td></td><td>Canadian Boundary Survey</td></tr>
<tr><td>1895–1897</td><td>Slight retreat</td><td>Bryant</td></tr>
<tr><td>1897–1899</td><td></td><td>Gilbert</td></tr>
<tr><td>1899–1901</td><td>Retreat west part 700–1000 ft.</td><td rowspan="2">Advance west part several hundred yards; retreat east part several hundred yards</td><td>U. S. Fish Commission</td></tr>
<tr><td>1901–1905</td><td></td><td>Tarr and Martin</td></tr>
<tr><td>1905–1906</td><td colspan="2">Continued advance west part; very little change east part</td><td>Tarr</td></tr>
<tr><td>1906–1909</td><td colspan="2">Continued advance of west part; slight advance and crevassing of east part</td><td>Tarr and Martin</td></tr>
<tr><td>1909–1910</td><td colspan="2">Continued advance of west part 660 feet; some retreat and some advance in east part, but slight increase in crevassing.</td><td>Martin</td></tr>
<tr><td>1910–1913</td><td colspan="2">Recession of western and southeastern margins</td><td>Martin</td></tr>
</table>

CHAPTER VII

THE VARIEGATED, ORANGE AND BUTLER GLACIERS

Variegated Glacier

Previous Description. Russell, the first scientific man to see this glacier which lies just east of Hubbard Glacier described it as follows [1] in 1890:—

"A débris-covered glacier, so completely concealed by continuous sheets of stones and earth that its true character can scarcely be recognized, descends from the mountains just east of Hubbard Glacier. It is formed by the union of two principal tributaries, and, on reaching comparatively level ground, expands into a broad ice foot, but does not have sufficient volume to reach the sea."

On his second expedition in 1891, having a better view of it, Russell decided that the Variegated Glacier coalesced with the Hubbard.[2]

"East of Hubbard Glacier there is a large buried glacier fed by several ice streams from the mountains above, which I judged from the view obtained from the *Corwin*, in 1890, was separated from the Hubbard Glacier, but better opportunities for observation obtained a year later showed that the two ice bodies are confluent."

Gilbert [3] reached the same conclusion in 1899, as did Gannett [4] in making his map. Gilbert's description of what we now call Variegated Glacier, based on seeing it from between Hubbard and Variegated Glaciers and upon a comprehensive view from the 1000 foot station above Osier Island, was as follows. He is writing about the combined Hubbard and Variegated Glaciers.

"The southeastern third of the glacier was moraine covered, not only at the water edge but for nearly or quite two miles inland. The material was coarse and angular, and was divided into zones or belts distinguished at a distance by their contrasting colors—black, yellow, purple, green, blue-black, and orange or rusty. These bands had not the ordinary arrangement of parallel medial moraines, but tended rather to contour the slope, and the search for their origin and meaning would make an interesting and profitable study. Some of them occupy ridges and others hollows, suggesting inequality in their ability to retard the melting of the ice beneath, but the whole surface was rugged in detail, exhibiting a continuous series of hummocks and kettles."

Our interpretation of the relationship of Hubbard and Variegated Glaciers, already stated in connection with Hubbard Glacier, is that, although the two glaciers undoubtedly coalesce, they are essentially independent (Map 3). The name Variegated has,

[1] Russell, I. C., An Expedition to Mount St. Elias, Nat. Geog. Mag., Vol. 3, 1891, p. 100.

[2] Russell, I. C., Second Expedition to Mount St. Elias, in 1891. Thirteenth Ann. Rept., U. S. Geol. Survey, pt. 2, 1892, p. 85.

[3] Gilbert, G. K., Glaciers and Glaciation, Harriman Alaska Expedition, Vol. 3, 1904, pp. 63–65.

[4] Gannett, Henry, Same, Plate VIII, opposite p. 62.

therefore been applied not only to the valley feeder, but also to the stagnant, piedmont area that spreads westward until it partly unites with the Hubbard.

In 1905 we assumed, with Russell and Gilbert, that the Variegated Glacier bulb was fed by two glaciers, the Variegated and Orange,[1] but in 1906, on a journey to the supposed eastern tributary, now named the Orange Glacier, it was found that, while this glacier did actually coalesce with the Variegated for a short distance, the two were quite independent, and the stagnant piedmont area was entirely supplied and dominated by the Variegated Glacier. It has therefore been necessary to change the names applied in 1905 under the supposition that the piedmont area was supplied by two arms.

The Variegated Glacier and its piedmont bulb presented many features of interest in 1905, especially the nature and distribution of its ablation moraines, but in 1906 it was absolutely transformed by a spasmodic advance. Therefore in both years it received considerable attention. The statement and discussion of the observations of 1905 and 1906, which are fully presented in the reports on the expedition of these years, are in the main so essential to a clear understanding of the changes observed in 1909, that it is necessary to restate the main points in the previously published description.

Valley Portion in 1905.[2] Variegated Glacier descends in serpentine course from an unknown source among the mountains east of Mt. Seattle through a deep valley bordered by mountain walls greatly steepened by glacial erosion. We ascended this glacier to an elevation of 2100 feet, finding the ice everywhere smooth and easy to traverse (Pl. LIII), with only here and there a crevasse, or small group of crevasses, in the upper portion, but nowhere finding our progress impeded. In cross-section the glacier surface was flat, with a descent on either margin, increasing in steepness toward the valley mouth. The glacier grade varied considerably, but within the mountains was usually from 7 to 10 degrees, though near the valley mouth the grade became much less, and in some areas the ice surface was quite flat, and even locally had an upstream descent. Below the snow line, which lay just above the highest point which we reached (2100 feet), ablation was rapidly in progress increasing greatly in the lower portion, where both lateral and medial moraines appeared, while scattered patches of moraine and angular blocks were strewn over the surface.

The evidence was complete that even well within its mountain valley the Variegated Glacier was only moderately active, while in the lower part of the valley portion the glacier was quite inactive. This inactivity of the lower portion was shown not only by the lack of crevassing, and the flat or undulating surface strewn with débris, but also by the presence of a marginal valley on either side with moraine-covered ice for one wall and the mountain side for the other. The smoothed and striated mountain walls, and the presence of moraine terraces on their slopes, on which only a few scattered annual plants had yet encroached, proved that at a very recent date this glacier had been more expanded and was then in contact with the mountain wall clear down to the end of the valley.

The moraine on the glacier was mainly angular and evidently largely derived from rock falls from the mountain sides, which include a variety of rock types, including black

[1] Tarr, R. S. and Martin, Lawrence, Glaciers and Glaciation of Yakutat Bay, Alaska, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 147–149.

[2] See Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 46–53.

hornblende gneiss near the valley mouth on both sides, and above this, green argillite schist, white granite, and red, orange, and banded-gray gneiss that on exposure to weather rusts to different colors. All these rock types were present in the angular moraine that was strewn in bands and patches over the lower portion of the valley glacier.

The Expanded Bulb in 1905. On emerging from its valley (Map 3, in pocket), Variegated Glacier expanded in a broad, bulb-shaped piedmont area, which was evidently completely stagnant. This bulb spread westward until it coalesced with Hubbard Glacier, and southward almost to the fiord. Throughout the piedmont bulb the surface was almost completely covered with ablation moraine, increasing in thickness toward the periphery.[1] The piedmont area consisted of three quite different parts: (1) just beyond the valley portion of the glacier an expanded bulb mostly covered by a series of crescentic moraine ridges of different colors,[2] though with ice appearing here and there in small areas; (2) a crescentic interior flat mainly of gravel, but with ice beneath; and (3) an outer, much larger area so deeply covered with moraine that ice showed in only a few places.

The Inner Bulb. Between the interior flat and the valley glacier the inner bulb began abruptly in a steep ice wall rising fully 100 feet above the partially débris-veneered portion of the valley glacier. This slope rose up to a ridge whose crest was covered with blocks of a gray quartz gneiss which on exposure to the weather rusted to an orange yellow, giving to the moraine ridge a distinct orange tinge noticeable at a distance of 7 or 8 miles. This orange moraine ridge was crescentic in form with the concave side upstream, and on its eastern end it merged into the eastern lateral moraine of the valley part of the glacier.

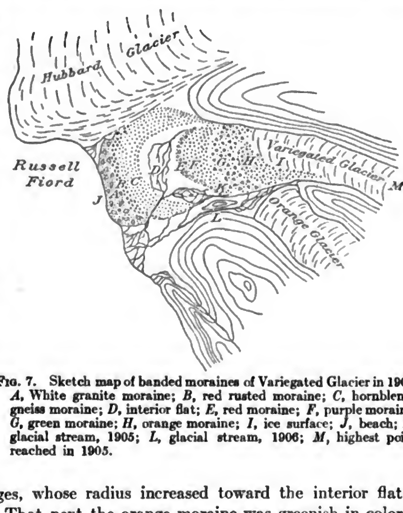


FIG. 7. Sketch map of banded moraines of Variegated Glacier in 1905. *A*, White granite moraine; *B*, red rusted moraine; *C*, hornblende gneiss moraine; *D*, interior flat; *E*, red moraine; *F*, purple moraine; *G*, green moraine; *H*, orange moraine; *I*, ice surface; *J*, beach; *K*, glacial stream, 1905; *L*, glacial stream, 1906; *M*, highest point reached in 1905.

Outside the orange moraine, down as far as the interior flat, were a succession of valleys and ridges, all crescentic in form (Fig. 7). The valleys were located in areas of thin moraine, while the ridges marked the sites of thicker morainic accumulation; but nowhere was the moraine thick. The crescentic ridges, whose radius increased toward the interior flat, were notably different in color. That next the orange moraine was greenish in color,

[1] For a more detailed description of this piedmont area, see Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 47–49.

[2] First observed and described by Gilbert in 1899.

due to a predominance of green schist, while beyond this came a purplish moraine, due to the presence of a purple gneiss in sufficient quantities to mask the scattered red, orange and black fragments in it. There were other shorter, less distinct moraines with curvature parallel to the neighboring more pronounced ridges. The purple moraine, which had a width of about an eighth of a mile, descended steeply to the interior flat, but on this descent were two minor, not very distinct ridges, one red in color, the other white, owing to the presence of many fragments of white granite.

Because of our interest in the peculiar banded moraines we traversed this part of the ice bulb and nowhere found the least evidence of motion. It was a greatly roughened surface, strewn with large angular blocks, but not a single crevasse was found. Here and there ice was seen, especially on the steeper ridge slopes and in the valleys, and everywhere there was clear evidence that ice existed just beneath the moraine surface, and we could usually reach it by thrusting our ice axes into the moraine. The surface was constantly shifting by the undermining caused by melting of the ice beneath, and probably this is the main reason why there was so little vegetation. Scattered willows, none over two years old, were growing here and there on all the moraine ridges excepting the orange moraine. Many dead plants were seen, and many others with their roots partly exposed by recent undermining. Evidently in such a shifting soil the growth of but a year or two was all that a plant could attain before the slumpingof the surface removed the soil in which it was growing.

The Interior Flat in 1905. The interior flat had the form of a crescent with the concave side toward the mountain valley, where it was bordered by the steeply-rising red and purple moraine whose crest was about 250 feet above it; on the outer side it was bordered by the concave ridge of the broad, black hornblende gneiss area of the outer bulb, which rose over 200 feet above it. The flat was, therefore, a very notable feature, (Pl. LI, A) interrupting the succession of crescentic moraines of the inner and outer bulbs. Its length was more than a mile and its width in its widest part fully three quarters of a mile, but narrowing rapidly toward either end. Its elevation was about 100 feet above sea level. The surface of the interior flat was mainly composed of gravels brought by streams issuing from the inner bulb of the Variegated Glacier, but some low ridges and hummocks rose above the alluvial flats. There were areas of depression occupied by lakes, some pits formed by recent settling, and numerous cracks and faults, proving that the settling was still in progress. These facts prove the presence of ice beneath the interior flat, but that it was not in motion was clearly indicated by the general levelness of the floor of the interior flat. Probably the ice beneath the flat was very thin.

It is believed that the interior flat represents the site of a clear ice area, such as is observed in other glaciers of this region, which, when the glacier became stagnant, was so lowered by ablation that it formed a depression in the glacier bulb in which sedimentation ultimately began, thus burying the ice. In 1905 two streams were contributing to the alluvial deposits, one from the northern side bringing the drainage of the northern margin of the Variegated Glacier, the other, and much the larger, emerging (Pl. LVII, A) from an ice gorge in the southern portion of the inner bulb, then descending in a fall into a granite gorge, from which it emerged first across moraine, then upon the interior flat. Both streams flowed in branching course across the interior flat, and then, after uniting, passed through a moraine-walled ice gorge in the outer bulb of the glaciers, and finally, uniting with a stream from Orange Glacier, flowing in many distributaries to the fiord over a very large

alluvial fan outside the periphery of the outer bulb. The ice drainage was therefore peculiar, for, on emerging from the glacier it passed first over a lowered part of the glacier which it was burying with its deposits, then through an ice gorge across the outer bulb, building another alluvial fan deposit there. With the development of the interior flat there has probably been a notable decrease in the rate of growth of the outer alluvial fan because so much of the sediment was left by the streams in the interior flat.

The Outer Bulb in 1905. The outer bulb, beyond the interior flat, had the same general features in 1905, 1906 and 1909. It rose steeply above the flat to a height of about a hundred feet, and thence to its periphery consisted of an irregular hummocky morainic mass. For the first half mile or so this moraine had a general black color due to predominance of black hornblende gneiss, though throughout its extent there were scattered fragments of white granite, purple gneiss and other rocks. Outside the black hornblende moraine came a narrow band of red rusted gneiss, then a belt a quarter of a mile or more in width, in which there was such an abundance of white granite fragments among the black hornblende as to give the moraine a grayish appearance when viewed from a distance. In the black hornblende belt there was everywhere distinct evidence of the presence of ice at no great distance beneath the surface, and every here and there ice was seen; but this moraine was much thicker than that on the inner bulb. No ice has been actually observed in the outer belt of white granite moraine, but we are convinced that ice existed there even to the very periphery where the bulb rises above the beach. We reach this conclusion because the moraine surface is cracked, there were areas of slumping, and much cold water was issuing in small streams or standing in shallow lakes. It is even possible that the ice extended out beneath the fiord, for there is here a broad shoal extending at least a quarter of a mile off shore.

On the black hornblende moraine there was a general absence of vegetation, though moss, tufts of grass, and scattered annual plants occurred here and there, and there were occasional young alders and willows, rarely more than one or two years old, though one was found with ten annual rings. Toward the periphery of the bulb, where the moraine was thicker, the ice thinner, and, therefore, the conditions of stability much greater, the amount of plant growth increased rapidly. On the very outer portion of the moraine were small alder thickets with bushes having usually no more than eight rings of annual growth.

It was evident in 1905 that no ice motion extended across the interior flat to the outer bulb, and, in fact, that there had been no motion in this outer portion of the glacier for a number of years before that. The outer bulb was, therefore, passing through a period of wastage with no addition to its supply, and the inner bulb had started upon the same career, having progressed far enough in it to become almost completely moraine-covered. The condition of 1905 gave reason for believing that both the outer and inner bulbs, together with the interior flat, were on the road to complete destruction by ablation, and had it not been for the spectacular changes of the succeeding year there would doubtless have been little of interest to record about this glacier except the progress of its wastage and the interpretation of its morainic phenomena.

The Advance of 1906. We saw Variegated Glacier for the last time in 1905 about the middle of August, and its condition was then as described above. About ten months later, in June, 1906, when the senior author approached it, he saw, even from a distance, that it was utterly transformed. The outer moraine, outside the interior flat was un-

changed, and the flat itself was unaltered excepting by being narrowed by advance of the ice on the inner side; but all above the interior flat was completely changed. Far up the valley, to the point where we had so easily walked ten months before, the ice surface was a maze of jagged pinnacles and seracs with intervening crevasses. The ice rose higher in its valley. It extended to the walls of the mountain valley. All smooth ice was gone, and the medial and lateral moraines were destroyed. It had become an entirely different glacier in almost every respect in the brief interval of ten months. From the valley mouth to its head the glacier was impassable, and there was not a half acre that was unbroken. The entire valley glacier was broken as an ice fall is in Alpine valleys, or as the Hubbard Glacier is back of its ice cliff. It seemed almost incredible that ten months before one could go over this surface in any direction, encountering a crevasse only here and there. In August, 1905, sleds could have been easily drawn over the valley glacier; in 1906 one could go out on the glacier only with the greatest difficulty.

The crevassing extended throughout practically the entire inner bulb, destroying the series of colored moraines of 1905, and replacing this area of moraine-covered ridges and hummocks with a maze of jagged pinnacles and deep crevasses. In 1905 from a distance no ice could be seen in this area of banded moraines, but in 1906 fully half the surface was clear ice. Much of the moraine had fallen into the crevasses, and the only indication of the former series of banded moraines was the presence, here and there, of a faint tinge of color on the broken ice surface. Our attempts to climb upon the broken ice to a point where we could photograph from approximately the same position which we occupied in 1905 were baffled by the intricate maze of crevasses which extended, in all directions, to the very edge of the inner bulb.

The periphery of the inner bulb had been pushed forward a 100 yards or more, overriding the granite gorge through which the glacial torrent emerged upon the interior flat in 1905 (Pl. LVII, A,) and narrowing the interior flat somewhat. There was a notable thickening of the inner bulb also, but we had no means of determining the amount, though the front cannot have been less than 200 or 300 feet higher in 1906 than it was in 1905. The advance of the front of the bulb upon the interior flat was in the nature of overriding rather than a pushing forward of the ice beneath it, plainly proving that the ice beneath the flat was very thin. For this reason, the effect of the thrust was not communicated across the flat to the outer bulb, which remained the same in 1906 as it was in the previous year.

One of the most notable changes brought about by the advance was upon the marginal drainage. By overriding the granite gorge the outflow of the glacial torrent at this point was checked and, for the time being, the interior flat was robbed of this, the greatest source for the alluvial deposits that had been accumulating in it. At the same time, naturally, the volume of the glacial stream that outflowed through the ice gorge across the outer bulb was correspondingly diminished and, therefore, the principal source of water for the building of the western part of the great alluvial fan beyond the outer bulb was withdrawn. This alluvial fan, the largest in Russell Fiord, was notable not only for its size but also because its eastern half was covered by an alder thicket, while its western half, crossed by the branches of the stream just mentioned, was barren of vegetation, as is normal for growing alluvial fans in this region. The presence of alder thickets on the eastern half of the fan was interpreted as proof of changes of earlier date. First there must have been streams flowing over it in order to build it; then, by some

change in the locus of emergence of drainage, the water was withdrawn from the western half of the fan allowing the development of an alder thicket. The maturity of the alders, indicates that the change in conditions occurred not less than a quarter of a century earlier. The cause for this change in drainage was inferred to be recession of the glacier, permitting a new point of outflow for the main drainage of the southern portion of the Variegated Glacier, and the natural inference was that this change was associated with the passage of the drainage through the granite gorge into the interior flat and thence out through the ice gorge upon the western margin of the alluvial fan.

The accuracy of this interpretation was verified in 1906, for then, by the closing of the granite gorge, the southern marginal drainage was forced to flow along the southern margin of the glacier and emerge upon the eastern side of the alluvial fan. This new locus of emergence turned the distributaries of the glacial torrent down over the eastern half of the alluvial fan, as it had doubtless done in earlier time, and the alder thicket was being inundated and destroyed. In one of our expeditions we were obliged to cross this part of the alluvial fan, fording dozens of muddy torrents, and we witnessed a stage in the destruction of the alder thicket from near at hand. The bushes were being submerged in ice-cold water and in the rapidly accumulating deposits of coarse gravel which the glacial torrent was unable to move further. Many of the bushes were uprooted and were being whirled off toward the sea; others were dying from the long-continued submergence beneath the ice waters; and others were being killed by the deep burial of their roots beneath the accumulating gravels. All the bushes were in full leaf, proving that the change had occurred during the growing season of 1906.

The Valley Portion in 1909. On our visit in the summer of 1909, we found that the advance of the Variegated Glacier had entirely ceased, and that here, too, ablation had healed the surface so that travel over it was again possible (Pl. LV). The surface was still very rough, and there were many crevasses, but it seemed probable that one could go up the glacier as far as we went in 1905, though with much greater difficulty.[1] We did not attempt this, though we went out to the center of the glacier, where it emerges from the mountains, and obtained a clear idea of the condition both in its upper and lower portions. One photograph, taken from a point not far away from the place where the red moraine stood in 1905, shows the difference in condition very clearly. In 1905 there was a smooth, dirt-stained glacier surface, terminating abruptly in a wall of red moraine which rose on the down-stream side (Pl. LIII). All below this was a rough, moraine-covered ice surface with the concentric bands of moraine already described; above, all was smooth ice with the exception of an occasional crevasse and a few small moraine ridges. In 1909, on the other hand, the clear ice area extended down much farther (Pl. LIV), the banded moraines were gone, and the ice surface consisted of a confused series of elevations and depressions with many shallow crevasses. We found it necessary to select our route with care where in 1905 we could go in almost any direction with equal ease. But, different as the surface was from the condition in 1905, it was even more strikingly different from its condition in 1906.

There are two other notable contrasts between the 1905 and 1909 condition of this glacier. In the former year there were marginal valleys where the glacier emerged from its mountain valley; but in 1909 these were gone and the ice was in contact with the

[1] In 1911 a Boundary Survey party went up Variegated Glacier to its very head and found no difficulty whatever in traveling over the surface which was impassably crevassed in 1906.

mountain wall without intervening depression. Further, the ice was perceptibly higher than in 1905. We could not determine the exact amount of thickening, though it was certainly not less than 100 feet where the glacier passes out from its mountain valley, but the ice surface was not so high as it was in 1906.

Banded Moraines on Expanded Bulb in 1909. The most noteworthy change in condition of the Variegated Glacier in 1909, as compared with 1905, was that the fine colored, banded moraines that in the former year covered the ice between the interior flat and the mountain valley were now gone. We could no longer trace the former succession. The purple and green moraines were entirely destroyed, and the other moraines were so disordered and displaced that we could no longer correlate them with the moraines of 1905; but ablation was at work covering the ice surface with new moraines of different kind and color (Fig. 8).

We crossed this glacier from the north side near the mouth of the mountain valley to the interior flat. There was abundant evidence that the glacier had been profoundly broken, but it was, nevertheless, a matter of unceasing astonishment that in so short an interval ablation had so effectually healed the broken glacier surface. Where the moraine cover was not thick enough to protect the ice, its surface was thrown into a series of great waves, 50 to 75 feet high, with the areas of greater crevassing mainly in the depressions. In addition, there were many smaller ice ridges and hummocks, all much rounded by ablation.

The area of clear ice extended far down below its former position, apparently a mile or more, and the pronounced red moraine that formerly rose as the lower border of the clear ice had apparently been carried bodily forward and incorporated in the ice. A small area along the northern margin, which was not profoundly affected by the advance, was still stagnant and retained a series of banded moraines; but everywhere else inside the interior flat the old series of variegated moraines, from which the name of the glacier was derived, had been destroyed. In their place an entirely new set of wholly different character and arrangement had begun to develop. On the southeast side a band of white granite moraine appeared, but it did not swing out into the ice in crescentic form. Then, after an intervening area of clear ice (Pl. LVI), a band of black hornblende moraine swept in a great crescent far across the ice bulb, which was here expanded beyond the mountain front. This morainic crescent was not complete, being most pronounced toward the sides and dying out completely near the center, where in its place was a red gneiss moraine much thinner and less pronounced than the one of black hornblende.

Both on its inner and outer face the black hornblende moraine rose about 100 feet above the clear ice. This gives a rough measure of the extent of ablation on this ice surface since 1906, for it is not to be supposed that the moraine had any notable initial elevation during the crevassing and forward motion of the ice. Although it is true that the ice under the ridge has also been somewhat lowered, it has been much protected by the débris, and in the meantime the clear ice has been lowered over 100 feet. Therefore, since 1906, the general ablation in this section has been in excess of 100 feet, and this fact gives a basis for understanding the remarkable effectiveness of ablation in smoothing the broken surface. One very noteworthy fact was the shallowness of the crevasses, for in 1909 the bottoms of even the deepest were visible. It was evidently only the rigid upper ice that was broken, and the breaking, though extending down over 100 feet, probably did not much exceed 200 feet.

Outside of the black moraine was a broad area of clear ice, which in time may give rise to a small interior flat, though possibly, as ablation proceeds, a sufficient amount of incorporated débris will be concentrated on the surface to prevent this. Beyond the area of clear ice, at the very edge of the broken portion of the glacier, and on its front, which rises above the interior flat, was a veneer of reddish orange moraine. This was apparently a mass of the old moraine cover pushed forward and mixed-up, at least such of it as has not been incorporated in the ice. All semblance of the former banding of moraine had disappeared and the only extensive morainic area in the broken portion of the glacier was this reddish orange moraine which covered the front of the inner bulb and a narrow strip back of it,—a total width of ¼ to ½ mile of moraine-veneered ice, where in 1905 there were two miles or more of banded moraine.

Excepting at the front of the bulb, near the interior flat, the morainic veneer was very

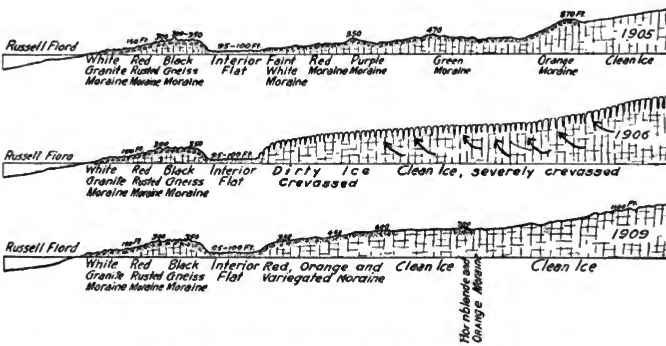


FIG. 8. LONGITUDINAL SECTIONS OF VARIEGATED GLACIER.

thin and unstable. Ablation was still rapidly in progress, even in the morainic area, but the débris cover was becoming thick enough and, over large areas, general enough so that the future rate of ablation will be materially checked.

Aside from the sudden advance, sudden cessation of the advance, and subsequent rapid ablation, one of the most notable features is the development of the black moraine (Fig. 8), apparently the result in part of the re-establishment of contact with the black hornblende portals to the mountain valley, in part to the outward movement of the black hornblende lateral moraine that before 1906 had developed in the marginal gully at the mouth of the mountain valley. In earlier days, when the glacier was in contact with these walls, its currents carried the débris out far enough to give rise to the great crescent that forms the outer wall of the interior flat next Hubbard Glacier; but by the spasmodic rush of 1906, which again established contact with the black hornblende slopes, the forward advance was so soon ended that the black moraine could not be carried out nearly so far. It lies a mile or more from the crescent on the outer margin of the flat. Had the ice advance continued, and the glacier spread out in a broader bulb,

and overridden the interior flat, the new black hornblende moraine might have been pushed out as far as the old one. One may infer, that the old black hornblende moraine was not the result of a spasmodic advance similar to that of 1906, but either of a much greater advance of this character, or else of a continuous advance maintained for a considerable period of time.

While the most striking feature of the Variegated Glacier in 1909, as compared with its condition in 1905, was the utter change in the moraines, there are other notable differences. One of these was the absence of well-defined ice drainage and of large moulins; but these were developing. Another noteworthy point is the evidence of lowering of the glacier surface since the advance of 1906. This has already been discussed in our consideration of the effect of ablation in smoothing the broken glacier surface. It is also clearly evident by a comparison of the earlier photographs with the 1909 condition. For instance, we reoccupied the sites of 1906 photographs on the south side of the glacier, and from there the lowering of the surface and change from a jagged surface to a rolling one was remarkable. Indeed, we estimate that it has amounted to at least 200 feet, and probably as much as 300 feet in the part of the glacier lying beyond the mountain valley, though less within this valley; but whether all this lowering is due to ablation is not certain. It is probable that there had been some settling due to further flowage of the glacier after the thickening which the spasmodic advance caused.

As compared with the Atrevida and Haenke Glaciers, while the main changes since 1906 have been the same, Variegated Glacier introduces one point of notable difference. The two other glaciers have reverted to their former completely moraine-veneered condition, the only exception being in the area of clear ice introduced in the Atrevida just outside the mountain front. But Variegated Glacier has not returned to its former condition, and the area of clear ice beyond the mountain front is greatly increased. A partial answer to the question as to the reason for this difference is undoubtedly that originally there was less thickness of ablation moraine on the Variegated bulb, this being possibly due to the greater stability of the crystalline rocks of its mountain valley as compared with the fissile and crumbling sedimentary rocks which enclose the valleys of the other two glaciers. But this answer is only partial, for the Variegated bulb was almost uniformly moraine-covered in 1905 and now it is not. This moraine has gonesomewhe re. It is not swallowed up in the crevasses, for the bottoms of most of them are visible; and the only conclusion that one can reach is that it has been carried forward. This, however, raises a new difficulty. Had the red moraine been bodily moved forward a mile, carried on the glacier surface while clear ice from above followed down behind it, the front should have moved forward approximately that amount; but it did not. To account for the increase in the area of clear ice, and the forward movement of the moraine cover without a pronounced advance of the glacier front, seems to us impossible of explanation on any other hypothesis than that clear ice from below has risen up here, and displaced and pushed the old moraine cover forward and to the sides, and that the transfer of ice and moraine has been accomplished by thickening of the bulb rather than by a pronounced forward movement of the front. With reference to the valley part of the glacier this clear ice area had the same relative position as that of the Atrevida. It is possible that in these two cases we have an illustration of the same phenomenon and an explanation of the development of clear ice areas that afterward form interior flats.

The Interior Flat in 1909. In 1909 we reoccupied essentially the same sites as those

of our 1905 and 1906 pictures, looking across the interior flat to the Variegated ice bulb. As compared with 1905, the front of the glacier where it rises above the interior flat was much higher, steeper, and more irregular; but as compared with 1906 it was both lower and much smoother. The serrated crest and steep glacier front, greatly broken by crevasses in which clear ice appeared in 1906, had by 1909 slumped into an undulating moraine-covered slope (Pl. LII) up which we could easily climb. We cannot tell exactly how much change had taken place, for it was impossible to occupy exactly the same photographic sites as in previous years, owing to slumping of the moraine from which our earlier pictures were taken; but there had certainly been a change amounting to scores of feet. Throughout this ice face it is evident that the morainic veneer is thin, excepting near the ice base to which much débris has fallen.

The most noteworthy fact with regard to the ice condition here is the evidence that when we observed the Variegated Glacier in August, 1906, it had advanced approximately to its maximum. Not only did the advance not affect the black hornblende moraine on the outer side of the interior flat, but it produced no further effect on the interior flat. Thus in an interval of less than a year an utter transformation of Variegated Glacier occurred, breaking its surface, thickening it, and causing some advance, but not affecting the outer stagnant portion. Since 1906 there had been no notable recession of the broken front, for willows grew right up to the base of the moraine; but recession would hardly be expected here, where moraine was still sliding down the steep front and forming a protecting talus at its base.

The granite gorge visible in 1905 was still covered by the glacier, but more water emerged from this portion than in 1906, though not nearly so much as in 1905. Evidently the subglacial drainage was again developing, and we may with safety predict a reoccupation of this area by a large glacial stream. There was no noteworthy difference in other parts of the interior flat and the description of its condition in 1905 and 1906 still applied to it. There was a fair-sized stream from the north side, apparently of about the same size as in previous years, but more water than in 1906 crossed the flat from the ice front and escaped through the gorge in the outer glacier to the sea.

The Outer Alluvial Fan in 1909. In 1909 we again visited the huge alluvial fan of the Orange and Variegated Glaciers, whose alder cover in 1906 was being destroyed by the floods of icy water and gravel that swept down the eastern half of the alluvial fan. The glacial torrent still emerged from the southern margin of the Variegated Glacier, apparently in nearly as great a volume as in 1906, but its course was now shifted to the central and western sides of the alluvial fan. Before its course was shifted, however, it had accomplished great destruction in the alder-covered zone. There was some alder left, notably a band on the eastern margin of the fan and a V-shaped wedge out toward the middle; but fully half the alder cover had been destroyed, and the alluvial fan surface was littered with uprooted alder bushes which had been washed down-stream and become stranded, some of them again taking root in the gravels in which they were embedded. Some of the alder still standing in place was partly buried by a deposit of a foot or more of coarse gravel, and some had been killed by deposit without being uprooted. Since we saw the alder still growing in the midst of the muddy distributaries in August, 1906, the season when the destruction began, and since much alder had been removed before 1909, we assume that the destruction continued throughout the season of 1907 and possibly into 1908; but in 1909, for a time at least, the destruction of the alder here had ceased.

The great volume of water involved in this work of destruction is clearly shown now by the network of small and large channel scars which extend across the formerly alder-covered part of the fan as well as by the score or two of streams which had to be forded in crossing the alluvial fan in 1906.

Banded Moraines Outside the Interior Flat in 1909. In general features the outer banded moraines of the Variegated Glacier bulb beyond the interior flat were in the same condition as in 1905. It is certain that the thrust from the advance of the Variegated Glacier did not affect this stagnant portion of its bulb in the least. In our excursions over it several observations in addition to those of previous years were made. Perhaps the most important of these was the clear indication of the presence of ice under even the outermost portion of the western edge, nearest Hubbard Glacier, and only a few hundred feet from the sea. While ice was not actually seen here, there was a small pond of cold water milky with sediment, such as could only come from ice, and on its shores were areas of fresh slumping clearly indicative of the presence of ice beneath.

Farther back from the fiord there was increasing evidence of the presence of ice, both in the amount of slumping and in the condition of the vegetation. Over most of the black hornblende moraine there were abundant small areas with no vegetation whatsoever. Then came little clusters of flowering plants, or equisetae, or grass, or willows from one to five years old; and now and then a large willow bush or even a cluster of them.

Marginal Drainage. The drainage condition is noteworthy. The northern marginal drainage followed the ice edge for a while, then cut across a part of the glacier in a narrow ice valley, then expanded over a broad interior flat, which rested on glacier ice and was completely bordered by ice except in the two gorges, one at the point where the drainage entered it, the other at the point where it left the interior flat. On this flat the stream was depositing much of its sediment; it then emerged from the interior flat through a gorge cut across the outer bulb of the glacier, and finally entered the sea over the large, partly alder-covered alluvial fan. In 1905 a still larger stream and perhaps practically the whole drainage of the Variegated Glacier and of Orange Glacier entered this flat and joined the north stream just above the outlet gorge. It is a most exceptional drainage condition, and the deposits left here when the ice melts out from beneath them will be complex and peculiar. There were no visible changes in drainage or moraine distribution by June, 1910, though time was not available for a re-examination that year.

Orange Glacier

General Description. Immediately east of Variegated Glacier is a broad valley occupied by a glacier somewhat more than a mile in width, which terminates close by the southern margin of Variegated Glacier in a low, gently-sloping ice front which on the northern side barely coalesces with the Variegated Glacier. From this front the Orange Glacier rises with moderate slope, varying from 5 to 15 degrees, though rarely attaining so steep a slope as the latter, finally reaching a broad, flat divide at an elevation of about 2500 feet. The divide area is above the snow line and from it the ice descends not only westward to form the Orange Glacier, but eastward towards Nunatak Fiord to form an unnamed glacier, to which reference will be made in the next section. The valley that this glacier occupies is therefore a through valley, and the glacier is a typical, though small example, of a through glacier.

The Orange Glacier surface is smooth and relatively free from crevassing, the few

crevasses nowhere interfering with free travel over the ice surface. From this it is inferred that the glacier is moving very slowly. Much of the surface is darkened by fine morainic débris, but, excepting at the margin and in the narrow bands in the middle, there is no continuous morainic cover. In fact, when viewed from a distance, the glacier has the appearance of being entirely free from moraine, with the exception of medial and lateral moraines. No large tributaries to this glacier are visible, and it is certain that none but small tributaries enter from the south, for on this side there is only a narrow range of mountains with peaks that rise but six or seven thousand feet. There is a possibility that several tributaries of medium size enter from the north, and one tributary is shown in a Boundary Survey photograph of 1906. There can hardly be any large tributaries from this side, because the Variegated and Butler Glaciers head back in the mountains to the north, and so near as to cut off any possibility of long feeders to the Orange Glacier. The slight motion in the Orange Glacier is further indication of the absence of large tributaries. It was, therefore, suggested in the report of the 1905 and 1906 expeditions that this was in reality a wasting through glacier, a remnant of a former greater through glacier which crossed this divide from Nunatak Fiord to Russell Fiord in the neighborhood of Hubbard Glacier. In our expedition of 1909 we saw no reason for questioning this hypothesis, nor did we discover any additional facts bearing upon it.

Contrast with Variegated Glacier. Orange Glacier is of special interest from two different standpoints, both of which may be made clear by contrast with the conditions in Variegated Glacier. The first of these contrasts is the difference in the general characteristics of the two glaciers. The Variegated Glacier winds down through a deep mountain valley, then, emerging from its valley, spreads out at the mountain base in a broad, expanded piedmont bulb. Orange Glacier, so far as we have seen it, is a broad, slowly-moving, through glacier spreading down in both directions from the divide area. This through glacier terminates on each end without any notable expansion and without any considerable area of ablation moraine. There also seems to be a distinct difference in the source of the ice of the two glaciers. That giving rise to the Variegated Glacier is derived from snow slides and numerous, short, tributary glaciers from the steeply-enclosed mountains. The ice supply for Orange Glacier is also partly derived from the sliding of snow and the descending of short tributaries from the enclosing mountains, but a considerable portion of the supply, and perhaps the major part of it, is the snow that accumulates on the broad, flat divide from which the ice descends eastward and westward.

The second noteworthy difference between the Orange and Variegated Glaciers is in their behavior during the period of observation. Variegated Glacier, was subjected to a spasmodic advance by which its surface was broken from as far up the mountain valley as we could see, far down into the stagnant piedmont bulb, but before 1909 it had ceased to advance. Orange Glacier, on the other hand, was in the same condition in 1906, as it was in 1905, and in 1909, 1910 and 1913 its condition had not changed. If this glacier is a wasting through glacier remnant there would be no reason to expect an advance in response to the earthquake shaking of 1899. If its ice supply comes mainly from the downsliding of snow and from short mountain tributaries, we should have expected some advance before 1910. The fact that there has been no such advance supports the interpretation of the origin of this glacier cited above. The only alternate hypothesis to account for the failure of this glacier to respond to the influence of earth-

quake shaking is that it is supplied by a long glacier from the north, which, flowing out upon the broad divide area, distributes its ice both to the east and to the west, and whose length is so great that the response to the earthquake shaking has not yet reached the two distributing arms. This does not seem probable to us.

Butler Glacier

General Description. Back of the low mountain range on the north side of Nunatak Fiord lies the east end of the Orange through glacier. This was not visited in 1905 or 1906 and the exact conditions were misinterpreted. It was then thought that the eastern end of the Orange through glacier expanded beyond the mountain front in a broad, moraine-covered piedmont ice bulb. In 1909 we discovered that this eastern end of the glacier actually terminates well within the mountain valley in a steeply-rising, moraine-covered ice front. This fact is further indication that the Orange through glacier is weak and is supplied from no very extensive sources. Otherwise it would necessarily extend at least to the end of its mountain valley. The eastern terminus of this glacier is evidently essentially stagnant, and its position is not materially different from what it was in 1905.

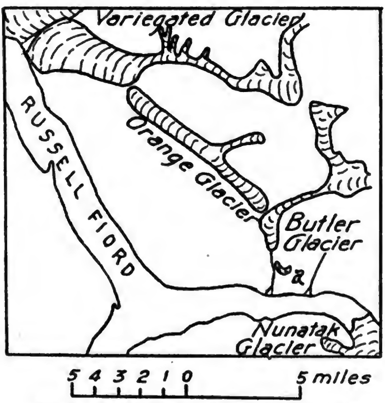


Fig. 9. Sketch of Butler and Orange Glaciers.

The moraine-covered piedmont ice bulb that was interpreted in 1905 and 1906 as an extension of the eastern end of the Orange through glacier was in 1909 found to be the piedmont bulb of Butler Glacier (Fig. 9), which descends from the mountains to the north. The Butler Glacier has not been studied within its mountain valley, and its conditions are, therefore, not known, but that it has considerable length, and a fairly-extensive supply, is indicated by the fact that it has been able to push itself out beyond the mountain front, whereas the much larger Orange Glacier has not been able to do so. In the view up the valley which we obtained we saw one good-sized hanging tributary and we also saw clear ice within the mountain valley.

The Butler Glacier Bulb. The part of Butler Glacier beyond the mountain front possesses features of distinct interest. It spreads out in a stagnant bulb completely across the Orange Glacier through valley, and is separated from the Orange Glacier by a well-defined depression, though it is possible that in this depression, which we did not visit, the two glaciers coalesce. From this expanded bulb a good-sized glacial stream emerges and flows through a broad, deep gorge (Pl. LVII, B) cut in well-assorted, stratified gravels which rise almost vertically to a height of fully 100 feet. Beyond this gorge

the stream divides into many branches over an alluvial fan which it is building out into the fiord. There is a large area of gravels to the southeast of the expanded ice and gorges cut in them show clearly that they form a thick deposit. One gorge, in particular, now no longer occupied by the stream, has in front of it a well-defined alluvial fan. This abandoned gorge and fan prove clearly that there have been important changes in the drainage along the front of the ice bulb of Butler Glacier. Even more noteworthy in this connection is the fact that the upper surface of the gravels is fluted, by glacial erosion, has a thin moraine veneer, and is crossed by stream channels. At two points there are well-defined stream-bed deposits in anastomasing channels, showing that at a previous stage glacial drainage cut across these gravels at an elevation of 100 feet above the present stream. This stage of drainage was probably at the time when the glacier extended further out to the heads of these channels which are now cut off by the development of the gorge cut in the gravels.

Between this part of the gravel terrace and the sea is a range of low, hummocky hills rising to an elevation of from 100 to 175 feet, which is undoubtedly a stagnant moraine-covered ice block (Pl. LI, B), entirely disconnected from the Orange Glacier, so far as we could see. In a part of this moraine nearest the glacier, and on the margin of the gorge cut by present-day drainage, buried ice was revealed in the cliff face. The section here from base to top is as follows: Just above the alluvial fan, stratified gravels rise to a height of 25 feet; above this are 50 feet of ice; and, on the top, till moraine 8 feet deep. The moraine-covered ice area extends down to within an eighth of a mile of the sea; but it grows lower toward the sea and encloses in a semi-circle an area where there is no ice. Throughout the surface of this moraine (a, Fig. 9) there is slumping of the soil, cracking of the ground, an abundant outflow of cold water, all clear indications of the presence of buried ice; and at one point, a few hundred feet south of the ice cliff already mentioned, ice was discovered by digging to a depth of about 2 feet at the base of the slumped surface. Even the outer face of the moraine has little vegetation and a tumbled appearance, indicative of recent sliding of the moraine through undermining. Over a large part of the surface of the moraine there is a moss growth; and willow bushes, 5 to 10 years old, are scattered over the entire area; but many are overturned by the sliding of the soil.

The interpretation of these conditions is not difficult. At some earlier stage a series of gravels were deposited here, which continue nearly up to Nunatak Glacier, everywhere showing evidence of overriding by the glacier, which, though it failed to remove them, fluted the surface into terrace form and left upon it a morainic veneer. These gravels, apparently contemporaneous with others in Russell Fiord, antedate the last great advance of the glaciers of the Yakutat Bay region when Nunatak Glacier extended through Nunatak Fiord and even far up Russell Fiord. When the Nunatak Glacier receded past the mouth of the Butler Glacier valley, Butler Glacier expanded in a broad piedmont ice bulb over the entire area between the mountain base and the fiord. This bulb became covered with ablation moraine over a large part of its surface, but a clear ice area existed between the present piedmont area and the stagnant moraine-covered area just described. This clear ice area has completely melted, and by reason of this fact, together with erosion by the glacial stream, the outer moraine-covered ice area is now completely separated from the glacier which gave it birth. We have here apparently the last stages in the

destruction of an interior flat area, similar to, though smaller than, that existing in the Variegated Glacier bulb.

In 1905 the Butler Glacier presented no evidence of recent advance that we recognized, though it is possible that it may have undergone a transformation similar to that of the Galiano Glacier. Such a change, followed by healing through ablation similar to that on the Galiano Glacier would not have been recognized by us in the absence of previous observations or photographs for comparison. There certainly has been no notable advance and breaking of this glacier between 1905 and 1910, for any distinctive change would have been recognizable in our photographs. We may, therefore, assume that unless this glacier passed through a cycle of advance due to the 1899 earthquakes prior to 1905, it has not yet entered upon it.

Observations in 1910 and 1913. In 1910 and 1913, the junior author made distant observations of Butler Glacier from near Nunatak Glacier. It was observed that there was an area of outwash gravels in the depression between the eastern end of Orange Glacier and the piedmont area of Butler Glacier.

As far as visible up its mountain valley Butler Glacier is clear of débris. It appeared crevassed in the valley portion in 1910 where no particular crevassing was observed in 1905 or 1909, but there was no increase in this area of crevassing from 1910 to 1913.

CHAPTER VIII

NUNATAK AND CASCADING GLACIERS

Nunatak Glacier

General Description.[1] As in the case of all the larger glaciers of this region the name was given by Russell. He saw this glacier in September, 1891, from Cape Enchantment, opposite the mouth of Nunatak Fiord and says[2] that "near where it enters the bay it is divided by a rounded butte of bare rock that rises through it like an island and that suggested the name 'Nunatak Glacier.'" A nunatak in Greenland is a mountain peak rising above the glacier, which surrounds it, and Russell speaks of this in another publication[3] as "a rounded dome of rock which rises through the ice and forms a nunatak." If the nunatak had this appearance in 1891 both of the tongues of the glacier must have extended much farther than they did in 1905; but from Russell's distant point of view he might well have thought the hill surrounded, even though the two tongues fell short of junction by a considerable amount. When mapped by the surveyors of the Canadian Boundary Commission, in 1895, the glacier terminated in two tongues, or distributaries, the northern and larger one ending in a sea cliff (Pls. LIX, LXVIII), the smaller, southern one on the land in a valley south of the nunatak. The glacier is still in this condition (Map 4, in pocket), though there has been material recession in each of the distributaries.

From the top of the nunatak one sees that at a distance of three or four miles from the sea two large tributaries unite to form the lower portion. The southern tributary heads upon a broad, flat, snow-covered divide, beyond which, according to the prospectors, the ice descends eastward toward the Alsek valley (Pl. LVIII). There is certainly an open valley without lofty mountains, in this direction. It is, therefore, a through glacier, and in 1898 it was traversed by prospectors as a highway to the Alsek. The Boundary Survey map indicates also a branch through a valley leading southward to the Hidden Glacier, but this is hidden from sight from the places where we obtained views of the glacier. We cannot tell whether the southern tributary of the Nunatak Glacier receives any notable tributaries, but the absence of medial moraine upon its surface indicates that there are no such tributaries. There is a well-defined lateral moraine on the south side, and another on the north side. The latter joins the south lateral moraine of the north arm and forms a prominent medial moraine on the lower glacier below the junction of the two tributaries. That the southern arm receives no notable contributions from tributaries is also indicated by its weakness, for the northern arm pushes over into its territory; it is failing to continue to supply the land tongue of the glacier with ice, and

[1] See Tarr, R. S. and Martin, Lawrence, Glaciers and Glaciation of Yakutat Bay, Alaska, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 149–150; Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 54–58.

[2] 13th Ann. Rept., U. S. Geol. Survey, pt. II, 1892, p. 86.

[3] Glaciers of North America, Boston, 1897, p. 95.

its surface is not greatly crevassed. The most crevassed part is in a marginal area just east of the nunatak where the conditions that cause crevassing will be explained later.

The northern arm, on the other hand, is very vigorous and is made by the union of at least three good-sized tributaries, and probably more. Two of these are visible from the crest of the nunatak, and a third is seen from the mountain side southwest of it. They are also indicated by the two prominent medial moraines which sweep down the central portion of this glacier, and the probability of other tributaries is inferred from the fact that far up the valley there are five medial moraine ribbons. Of the three large visible tributaries which combine to make this arm the southernmost is apparently a through glacier, like the south arm, leading over to the Alsek valley, across a broad, flat, snow-covered divide from which the ice flows in both directions. The central tributary of the north arm comes down as a winding valley glacier of large size from among a distant group of high mountains that lie to the northeast. From the sea to the head of this valley cannot be less than 15 or 20 miles. The northern tributary, which is apparently much the largest and most active of the three, comes from the northwest, and all but the very lower portion is hidden from view behind the mountain wall of the north side of the Nunatak valley. There are some very lofty mountains here, and doubtless numerous tributaries descend from them. This arm of the Nunatak Glacier reaches back to the sources of the Hubbard Glacier as a through glacier.

Thus, while we do not know even the general features of the upper part of the Nunatak Glacier, we do know that it is a long glacier fed by several large contributing glaciers, and that it must rank as one of the largest glaciers of the region. That the glacier is actively moving is inferred from its profoundly-crevassed condition from its sea face (Pl. LIX) as far up the northern arm and its tributaries as we can see. It is, however, much smaller and less active than the Hubbard Glacier. The surface slope is moderate and the snow line fully ten miles from the sea, so that a very large area is exposed to ablation, but not enough to cover the lower end of the sea tongue with moraine. The snow line seems to lie higher on this glacier than on the Hubbard, probably because of the sweep which the ocean winds have across the two broad gaps leading toward the Alsek valley.

There is a pronounced lateral moraine on the north side of the northern arm, and another on the south side just above the junction of the north and south arms, which contributes to the medial moraine already mentioned. There are also two winding medial moraines which are pushed first to the east by the incoming of the northwestern tributary, then to the west by the incoming of the eastern tributary; then, after flowing medially down nearly to the junction of the two arms, these moraines are again turned at right angles toward the west and thence they extend to the sea (Pl. LX, A) near and parallel to the medial moraine formed by the junction of the north and south arms. The two medial moraines of the north arm are gray in color, while that made by the union of the two arms is black, suggesting that the mountain at whose base it is formed is made of black hornblende gneiss. That the northern arm is far the stronger of the two is certain from the course pursued by the three medial moraines, which show the ice currents. While they are turning westward they are also extending out across the southern arm; and the medial moraine made by the union of the two arms reaches a point nearly opposite the center of this arm before it is completely turned. It enters the sea in the southern quarter of the glacier. We, therefore, infer that, at most, the south arm of

the Nunatak Glacier supplies ice only for the land tongue and for not more than a quarter of the sea tongue.

Below the union of the two arms the united trunk glacier flows on for two miles or so, then splits on the east end of the nunatak into two distributaries, the land tongue and the sea tongue. The nunatak against which the glacier splits is a double-crested gneiss hill about 2 miles long and 1440 feet high at the highest point. It is extended westward by a low spur, or tail, of weaker schist and slate with the fiord on one side and a shallow ice-eroded valley on the other. The eastern end of the nunatak, which is somewhat lower than the centre of the hill, gradually narrows and finally descends beneath the glacier; but a crevassing of the ice in this direction may represent subglacial continuation of this hill eastward. Thus the nunatak and its ice-covered extension protect the land terminus from the thrust of the powerful north arm of the glacier, and tend to confine its supply to the south arm.

The Land Tongue, or Smaller Distributary. Gilbert photographed this tongue from two points, and Gannett mapped it as a distributary of the Nunatak Glacier. But in his description of the glacier Gilbert says [1] "In the hollow separating this knob (the nunatak) from the south wall lay a mass of ice of uncertain relations. It was seen only from the west, and was supposed to be a tongue or distributary arm of Nunatak Glacier. The fact that it lay several hundred feet higher than the tidal arm has raised doubts as to the correctness of the first impression, and I now suspect that it was only the remnant of a former arm of the glacier, stranded as a motionless and slowly-wasting summit mass. On the map of the Canadian Boundary Commission (1895) it is represented as a distributary of the glacier." Our observations and photographs prove clearly that it is a distributary, branching from the main glacier (Pl. LXI, A) at a sufficient elevation to permit of its extension down a valley higher than that of the sea tongue, and its termination at a level above that of the tidal end.

This distributary of Nunatak Glacier has suffered very material change since it was described and photographed by Gilbert in 1899. In 1905 the ice surface was smooth, with a slope of about 5°, increasing to 10° at the front, and there was no notable crevassing. On the nunatak side there was a small but well-defined lateral moraine, but on the southern side the lateral moraine was very broad and irregular and rose in places 75 feet above the glacier surface. It varied greatly in breadth, being broadest where large amounts of débris had fallen from the mountain wall, and particularly just where the distributary leaves the main valley. The fact that this fallen débris was not moved on to form a more regular band of lateral moraine testifies to the stagnant condition of this ice tongue. For a considerable distance from the valley wall the glacier was hidden beneath this moraine, which, though only a thin veneer, so irregularly protected the under ice that it had a hummocky topography. Recent extension of the glacier front westward was proved by the presence of this moraine-covered ice out beneath the Cascading Glacier. In 1909 ice still existed beneath this lateral moraine, which projected beyond the glacier front. The ice was protected from melting by the moraine which covers it and by the ice block talus that has fallen upon it from the Cascading Glacier, which is perched on the mountain slope above it.

Between the lateral moraines the ice surface was smooth and uncrevassed and evidently stagnant. The ice was rapidly melting, and numerous short streams coursed over

[1] Glaciers and Glaciation, Harriman Alaska Expedition, Vol. 3, 1904, pp. 61–62.

it and fell into moulins. The lower half of this distributary was stained with a thin film of débris, increasing in amount toward the end. The film formed a general sheet of clay and sand with some pebbles and bowlders, but was not thick enough to completely hide the ice. The finer material was being carried off in the streams. Here and there were pyramids of débris, with ice cores, varying in height from an inch to 6 or 8 feet, evidently on the site of earlier depressions in which the débris had gathered to later be etched into relief because of the protection against ablation which they gave to these parts of the wasting glacier surface. There were also numerous short ridges of similar character marking the sites of former crevasses. The outer edge of the glacier thinned down so that we could not determine its exact position. This edge was not covered by ablation moraine, but in places was buried beneath stratified deposits laid down by the water which escaped from the melting glacier. Some of these deposits were laid down in a temporary marginal lake, some as alluvial fan deposits; and all were laid down since 1899, for they occupied the site of the glacier end as photographed by Gilbert in that year.

A good sized stream emerged from the glacier front and flowed through the depression in which the stratified deposits lay, then entered a deep gorge in the friable slates and schists, which turns at right angles across the tail of the nunatak (Fig. 10; Pl. LXIII). Gilbert's photographs prove that in 1899 the glacier front extended into this gorge. After emerging from the gorge the stream spreads out in numerous branches over an alluvial fan which it is building in Nunatak Fiord, at the western base of the nunatak. We have been puzzled to explain this gorge for, from Russell's description of the condition of Nunatak Glacier, it seems certain that, even though the two arms of the glacier may not have been actually united in 1891, the land arm must surely have covered at least the upper part of the gorge in order to give him the impression that the hill was a nunatak. The gorge is certainly too broad and deep for the work of such a glacial stream in the 14 years between 1891 and 1905. It may of course be the product of stream work before the last advance of the Nunatak Glacier, or it may in part have resulted from subglacial stream erosion when the land tongue extended down beyond the western end of the nunatak.

As compared with the condition in 1899 the land arm of Nunatak Glacier in 1905 was notably different in several respects. Both lateral moraine areas had broadened considerably, thus reducing the area of clear ice. At the same time, the extent of débris-stained ice between the lateral moraines had increased, evidently as a result of the lowering of the glacier surface, which was very noticeable by comparison with Gilbert's photographs. The glacier surface of 1905 was certainly as much as 100 feet lower than it was in 1899, and its front had receded not less than two or three hundred yards.

In 1909 we found evidence of the continuation of the same changes (Pl. LXI, B). There had not been a very great recession of the front, although there had been some, but the glacier surface was very much lower and the lateral moraine areas were still broader and more in relief than in 1905. The very outermost edge of the glacier was a low, flat sheet of débris-stained ice (Pl. LXII, A). This distributary is evidently a dying end, no longer in motion, and, unless an advance begins, its rate of recession and period of ultimate destruction will depend entirely on the rate at which ablation can remove the ice; and this rate, though not known quantitatively, is certainly very rapid. A few decades should destroy this tongue if no advance comes in the meantime.

There is evidence that it has been steadily receding for a long time. Far out beyond

the position of its 1899 terminus the hill and valley surface consists of patches of bare, grooved, and polished rocks which frost action has not had time to roughen greatly, alternating with gravel and moraine areas in which the vegetation is young. It cannot have been many decades since this tongue pushed out far enough to have united with the tidal arm of Nunatak Glacier. It is possible, though not probable, that it did so in 1891 and it is unfortunate that Russell did not obtain near enough views of this glacier for us to know its position with certainty. The Canadian Boundary survey photographs show the hill as a semi-nunatak in 1895, with the land tongue occupying the head of the gorge. The first photograph showing its position from near at hand was taken by Gilbert in 1899, but we found in 1909 and 1910 some evidence for the preceding year. In 1898 some 300 prospectors passed over the Nunatak Glacier highway to the Alsek valley, and some returned by this route. On their return they abandoned their outfits, presumably as soon as they were useless and at the point farthest down the glacier over which they could draw their sledges. Roughly, with an error of only a few yards, the places where we found these sledges and other relics would represent the position of the ice front in 1898. We found a half dozen sledges, several rusty Yukon stoves, some ice creepers, spiked boots, etc. (Pl. LXII, B), at a point a little less than a half mile west of the 1909 and 1910 ice fronts and just south of where Gilbert's photographs show the terminus to have been in 1899. Aside from the improbability that men would draw sledges over bare rock (in the fall) and up hill and beyond the glacier terminus, the fact that the outfits were not being taken beyond where the ice ended in 1898 is established by the presence of an impassable stream gorge between the positions of the sledges, which were in two groups, and the present glacier terminus. It is, therefore, concluded that the ice ended in 1898 about where the sledges were found and that the retreat of the land distributary between 1898 and 1909 has been a little less than a half mile.

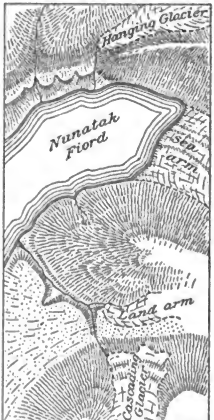


Fig. 10. Nunatak Glacier in 1905, Showing Gorge near Land-ending Arm.

In each of our five years of observation here, 1905, 1906, 1909, 1910 and 1913, we carefully observed the land tongue of Nunatak Glacier but without finding any evidence of advance. Being a glacier distributary still connected with an active glacier, but itself stagnant, it would be very quick to show response to even moderate advance. The fact that we have not detected any advance, together with the evidence that there was pronounced recession between 1899 and 1905, is accepted by us as definite proof that as yet there has been no advance as a result of the earthquakes of 1899. In 1910 and 1913, the junior author observed a continuation of retreat of the land arm of Nunatak Glacier which was then not essentially different from the 1909 condition except in the thinness of the terminal feather edge. The moraine to the right of it seemed to stand slightly higher in relief in 1910 than the year before, as comparative photographs showed.

This south distributary was traversed eastward to the very end of the nunatak to make sure of the contact with the tidal distributary, which changed so much between 1909 and 1910. The snow line was very low at the time of this journey, June 17, 1910, reaching practically to the end of the land arm. Nevertheless long strips of glacier surface had been bared by the wind, and the snow was nowhere very deep, so it was clearly evident throughout the journey that there had been no change in this arm of the glacier since 1905 and 1909, except lowering by ablation. The hard coarsely-crystalline ice was not at all crevassed.

From the extreme end of the nunatak, which we had not visited before, it was plain that the area of crevassing that extends southeastward from the nunatak is of a somewhat different nature than more distant views had led us to infer. It is not an elongated dome of crevassing which we have thought was wholly due to the exposure of the continuation of the rock ridge of the nunatak. Instead it is a crevassed northeast edge of the more stagnant arm of the glacier and in 1910 stood several hundred feet higher than the crevassed surface of the tidal arm which flows to the fiord. The crevasses are in gigantic steps descending northeastward and breaking the edge of the stagnant arm. Evidently during the two and a half mile retreat of the tidal distributary, between 1891 and 1909, its surface has been lowered faster by ablation and flowage combined than the stagnant arm, where ablation works alone and where the retreat was only a half mile or so. The north edge of the stagnant portion is, therefore, breaking because of removal of support and is perhaps preparing to flow later into the tidal distributary instead of the smaller arm south of the nunatak, down which it discharged in former times. The rock ledge continuing the nunatak is doubtless here as we have hitherto inferred, but the lower ice on the north side seems primarily responsible for permitting the crevassing.

The Sea Tongue, or Tidal Distributary. The tidal distributary of Nunatak Glacier, which is far larger than the land tongue, though not now extending so far west, is greatly crevassed from side to side. When visited by us in 1905 and 1906 the sea face was less than a mile wide, ending in a deep fiord bordered by precipitous mountain walls against which rested narrow lateral moraines which projected slightly farther than the ice front, especially in the case of the narrower south moraine. There were three medial moraines and almost no other débris on the glacier. The ice front, which rose precipitously for over 200 feet, projected farthest in the middle, and discharged icebergs almost constantly, the iceberg waves giving rise to violent surf which was doing important wave-erosion work on several miles of adjacent coast. The icebergs, mostly free from débris and including masses that rose 20 to 30 feet out of the water, were less numerous than those from Hubbard and Turner Glaciers.

Two streams emerged from the glacier below sea level, each about 200 yards from the glacier margin, giving rise to swirling currents of muddy water which kept floating ice away from the sea cliff, except just after an iceberg fall. The northern stream, which was the larger, showed its influence for over a mile. Between the streams there was a mass of floating ice close to the glacier front. These streams were thought to be accumulating a large deposit of sediment beneath the waters of the fiord.

Recession of the Tidal Distributary. As in the case of the land tongue of Nunatak Glacier, the tidal distributary furnishes clear evidence of notable recent recession. Russell's description in 1891 indicates that the glacier was much farther out, and the general absence of vegetation on the margin of the fiord, as well as the freshly-smoothed

and polished rock surfaces, both on the nunatak and on the mountain slopes, are convincing evidence of recent great recession. The first definite evidence as to the position of the ice front is furnished by the Canadian Boundary Commission map, based upon photographs made in 1895. In June, 1899, Gilbert made photographs of the tidal end of the Nunatak Glacier, and Gannett made a map showing its position (Pl. LXIX, B). From Gilbert's photographs and Gannett's map, compared with the Boundary Commission map, it is evident that between 1895 and 1899 there was a recession of the glacier front of a little less than a mile. There was an even more notable retreat between 1895 and 1905.

In 1905 we were able to reoccupy the exact site of Gilbert's 1899 photograph of the tidal end of Nunatak Glacier (Sta. A, Map 4), even to the extent of locating small bowlders in the immediate foreground. This 1899 photograph is reproduced here as Pl. LXIV, together with our 1905, 1906 and 1909 photographs from the same site (Pl. LXV). The amount of thinning and recession in six years is striking, the retreat being a little over a mile. We thought then that the great, and for this region exceptional, recession of this glacier front was possibly to be attributed to a shattering of the glacier by the earthquakes of September, 1899; but in view of the fact that the recession went on at least as fast for the four years before the earthquakes, we are now inclined to believe that this was not the case.

We occupied the Gilbert site again in 1906, and at that time the glacier front just barely appeared from behind the projecting nunatak and was much thinner. It then became evident that if recession continued this site would no longer possess importance because of the recession of the glacier out of sight behind the nunatak. This we found to be the case in 1909, when we again occupied the photographic site, and, as our photograph shows (Pl. LXV, C), no ice appears. The total recession of the glacier between 1905 and 1909 was about five-eighths of a mile, making a recession between 1899 and 1909, of a little over a mile and a half, or at the rate of nearly 860 feet a year. That this recession is exceptionally rapid, and unexpected by Dr. Gilbert, is indicated by the fact that he chose the photographic site he did. One could scarcely have predicted such an extremely great recession of the glacier front in so short a time.

In 1905 and 1906 we photographed the Nunatak Glacier from other viewpoints so that in 1909 we had a basis for noting the recession, which could no longer be seen from the original Gilbert site. One of these photographic sites which we reoccupied was on the crest of the nunatak, looking down upon the glacier front. From here we found that the glacier had receded greatly, and the outline of its front had changed completely. By examination of the distinctive small gullies (as b, Pl. LXVI) cut in the gravels that rest on the steep mountain slope on the northern side of the fiord, we are able to tell exactly where the ice front stood in the two years (a, ice edge in 1905, Pl. LXVI, A; c, ice edge in 1909, Pl. LXVI, B), and it is from this site that we estimate the amount of recession since 1905, which was about half a mile. With the recession of the glaciers there has apparently been a change in the position of outflow of the subglacial streams. In 1905 and 1906 there were two such streams, the larger emerging on the north side, the smaller on the south. In 1909 the largest stream came out from beneath the glacier near the south margin, while there was a smaller one in the middle, and another small one on the north. Another notable change in 1909, as compared with previous years, was in the discharge of icebergs. Formerly icebergs were so frequently discharged that one

needed to wait only a short time to witness large ice falls from the glacier front; but in our visit of 1909 we rarely saw or heard such ice falls, and the fiord in front of the glacier was far freer from floating ice than in former years.

Viewed from our photographic sites on the north side of Nunatak Fiord, the Nunatak Glacier presented even more notable changes. In 1909 the width of the glacier at the terminus was only $\frac{3}{4}$ of a mile and the marginal projections noted in 1905 were entirely gone (Pl. LXVII, B). The middle projection was also destroyed and the glacier front extended in an almost straight line from the north side to a point two-thirds of the way across the fiord where there was a very slight projection, beyond which the ice front receded. There was also a striking difference in position of the three medial moraines.

In 1905 (Pl. LXVII, A) these moraines entered the sea near the middle of the glacier. In 1909 they were all over in the southern half of the glacier. The black medial moraine caused by the union of the two arms of the Nunatak Glacier entered the sea in the southern quarter of the glacier front. The reason for this change in position of the medial moraines is explained by the appearance of a bulging of the ice surface across the southern arm where the nunatak hill extends beneath the glacier. It was evident that the southern tributary had been so diminished in activity that its surface had been lowered by ablation until this subglacial rock barrier became effective enough to cut off a considerable part of the supply that the southern arm formerly contributed to the tidal end of Nunatak Glacier. With this diminution in supply from the southern tributary, the still vigorous northern arm was able to thrust the medial moraines farther and farther toward the south.

Another noteworthy evidence of recession was clearly brought out from the photographic sites which were established on the alluvial fans on the north side of the fiord a short distance from sea level. From one viewpoint (Sta. F, Map 4) the front of the ice cliff was so near in 1905 that it hid the glacier so that we could not see much of its surface. Possibly also the ice cliff was higher than now. In 1909, on the other hand, from these same positions we were able to look up on the ice surface for a considerable distance. From this point of view it was seen in 1909 that back of its ice cliff the glacier rose in a series of steps; and the rise was so rapid in the lower mile that it seemed certain that not much more recession would be needed to transform this tidal glacier to one resting on the land. At the rate of recession going on in 1909 it would have been surprising if the Nunatak Glacier had been still tidal at the end of another ten years or even less.

The rapid recession of Nunatak Glacier between 1891 and 1909 seems to be part of a recession of much greater extent, perhaps continuous, perhaps interrupted by halts and readvances, but certainly never interrupted by any long period of halting. The evidence of the former extension of Nunatak Glacier until it coalesced with Hidden Glacier, and reached far up Russell Fiord, has been stated in the reports on our 1905 and 1906 expeditions, and will not be repeated here. That the complete recession from this former extension of the glacier has occurred in a brief time, is proved by the condition of vegetation which is now advancing to reoccupy the area from which the glacier has so recently receded. Positive and definite evidence of the amount of recession for a portion of the distance has just been stated in comparison of the photographic records of 1895, 1899, 1905, 1906, 1909 and 1910. The 1899 photographs by Gilbert show clearly the position of the ice edge in that year, and in 1905, 1909 and 1910 some of the ice still remained in a stagnant, débris-covered mass, detached from the main glacier and lying in about the position occupied in 1899 by the outermost moraine-covered northern edge of

the glacier. Off shore from this point, in 1909, there was a submerged reef, extending three-quarters of a mile from shore, sufficiently shallow for the stranding of large icebergs. Soundings made in 1910 show that this is probably part of the morainic deposit made during a halt in the glacial recession. A halt is suggested by the comparison between the position of Nunatak Glacier in 1895, as indicated by the Boundary Commission map, and its known position in 1899. Upon this evidence the rate of recession between 1895 and 1899 was about 1150 feet a year, or half as fast again as between 1899 and 1909. If recession between 1891 and 1895 was as fast even as the rate between 1899 and 1909 (860 feet per year) it would carry the tidal arm of the glacier out far enough to have led Russell, from his distant point of view at Cape Enchantment, to have inferred that the land tongue and the extended tidal end of Nunatak Glacier actually coalesced. It is to be remembered, however, that Russell did not specifically state that the nunatak was actually surrounded in 1891, though the name "Nunatak," and the statement that it "rises through" the glacier "like an island," warrant this interpretation of his description. The rate of recession may have been even more rapid between 1891 and 1895, for the rate from 1895 to 1899 was apparently one and a half times as fast as the rate from 1899 to 1909, perhaps partly as a result of the change in exposure of the glacier front, from southwest to northwest, with the bending of the fiord, and partly of the retreat of the glacier to a position in a narrower fiord with the high nunatak rising south of the 1899 to 1909 ice front, whereas at earlier stages the southern side of the broader glacier rested against low ground which separates the fiord from the southern mountain wall.

Our observations on the tidal distributary of Nunatak Glacier corroborate the conclusion reached from our observations on the land distributary that up to 1909 there was no response in this glacier from the effect of the earthquake shaking of 1899. Nunatak Glacier had undergone steady and extensive recession during the period of observation (Pl. LXIX). In this respect it contrasts with the two other tidal glaciers of Yakutat Bay, the Turner and Hubbard, both of which give indication of minor fluctuations in their ice fronts. These two glaciers have either maintained their position or have actually advanced, as has the other large glacier, the Hidden. We felt after our observations in 1909 that it could hardly be believed that in this region, where so many glaciers have advanced under the influence of the 1899 earthquakes, a large glacier like the Nunatak would fail to ultimately respond. We therefore anticipated that before many years the period of recession, so long and effectively in progress, would be brought to an end and be replaced by a sudden forward movement.

Advance of the Tidal Distributary in 1910. The junior author found that between July 24, 1909, and June 17, 1910, the retreat of the tidal arm of Nunatak Glacier had ceased and an advance begun. Our resurvey of the ice front plotted accurately in 1910 on a plane table map made a little less than eleven months before (Map 4), proved the advance to have varied in different parts of the ice front, the advance of the north side of the glacier being 1000 feet, the south side 350 to 400 feet, and the middle of the glacier 700 feet. The 1910 ice front is shown with that of 1909 on Plate LXX. Both margins of the glacier were seen to have extended northwestward down the fiord and the advance of the middle was shown by the lengthening of the medial moraines.

This is well shown in Pl. LXX where the photographs in 1909 and 1910 show the advance on both the left (north side of fiord) and on the right margin, as well as an

increase in curvature of the three prominent medial moraines and their extension to the right with the advance of the glacier down the fiord. These pictures also suggest an explanation of the lower bend in these medials in connection with the crowding down of the unsupported ice edge of the stagnant south tributary.

Comparison of the 1909 and 1910 photographs shows clearly that the 1910 ice front was an intermediate position between that of 1905 and that of 1909, the margins which receded during this period of four years of retreat not having yet advanced in 1910 to the 1905 position. The lengthening of the medial moraines was also apparent and there seemed to be a thickening on the north side near the curve in the crevassed north tributary. By comparing these two photographs with the 1905 picture from the same site it was evident that at the east end of the nunatak the crevassing which extends into the edge of the stagnant south tributary was more extensive in 1909 than in 1905 and less extensive in 1910 than in 1909, as would be natural if the main crevassed glacier was thinned by ablation between 1905 and 1909 and thickened again by advance between 1909 and 1910.

Soundings in the fiord in 1910, discussed more fully in Chapter XI, showed that the depth of water about a thousand feet west of the ice front of Nunatak Glacier was 555

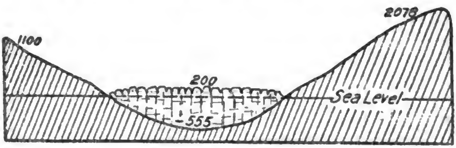


FIG. 11. CROSS-SECTION OF NUNATAK GLACIER AND FIORD. VERTICAL AND HORIZONTAL SCALE THE SAME.

feet. A true scale cross-section of Nunatak Glacier when it was at that point (sometime between 1906 and 1909) is reproduced as Fig. 11 and shows (a) that the glacier was about 750 feet thick; (b) that as the portion above sea level was only 200 feet (two-sevenths of the thickness), the glacier could not possibly be afloat; (c) that the slopes of the fiord walls above and below sea level are not significantly different; (d) that the proportion of the glacial valley now occupied by the ice is much less than when the greater glacier overrode and rose high above the nunatak.

Nunatak Glacier from 1911 to 1913. In 1911 and 1912 the Nunatak Glacier was visited by a Boundary Survey party under the direction of N. J. Ogilvie. In 1911 they observed a slight continuation of the advance of the previous year. In 1912 the glacier had commenced to melt back again, having retreated an amount estimated to be a quarter mile.

In September, 1913, the junior author revisited the Nunatak Glacier with the International Geological Congress. It was evident that the tidal ice front was still retreating, the northern margin projecting farther west than the center of the glacier, but no precise measurements of the amount of diminution were obtained.

Relation to Earthquake. The advance which commenced in 1910 was a minor one, such as has affected Turner and Hubbard Glaciers several times since 1899, not a greater one, such as has come about in Galiano, Haenke, Atrevida, Lucia, Marvine, and

Hidden Glaciers. In a tidal ice tongue like the Nunatak Glacier the future behavior is to be watched with the utmost interest. If the advance is attributable to the avalanche supply during the 1899 earthquakes, as we believe, this glacier has taken between ten and eleven years to respond to the accession in the snowfields and upper glacial tributaries. If the slight advance in 1910 was due to the thrust of a small tributary there may yet come a greater advance due to the activity of larger tributaries. If no advance comes in the next few years we shall perhaps be warranted in ascribing the lack of great advances of the Nunatak, Hubbard, and Turner Glaciers to their tidal condition.

The following tabular statement shows the known history of the tidal arm of Nunatak Glacier up to 1913:

Year	*Nature of Change*	*Observer*
Before 1891	Probably retreat	Russell
1891–1895	Retreat	Boundary Survey
1895–1899	Retreat about 4600 feet	Gilbert
1899–1905	Retreat about 5300 feet	Tarr and Martin
1905–1906	Retreat	Tarr
1905–1909	Retreat about 3300 feet	Tarr and Martin
1909–1910	Advance 700–1000 feet	Martin
1910–1911	Advance, slight	Boundary Survey
1911–1912	Retreat about ¼ mile	Boundary Survey
1912–1913	Retreat, slight	Martin

Cascading Glacier

General Description. Gilbert[1] who photographed and named the Cascading Glacier in 1899, when Gannett[2] showed it upon the map of Nunatak Fiord, speaks of it as occupying a high valley nearly at right angles to the Nunatak trough. "It was seen only as a series of ice cascades, pouring from ledge to ledge for a thousand feet down the steep wall of the trough." Its valley lies entirely above snow line and heads a few miles to the south in the mountains that lie between Hidden Glacier and Nunatak Fiord. The characteristics of the glacier in this valley are unknown excepting at the very edge, but it is probably not materially unlike scores of other short, high, mountain valley glaciers.

The most noteworthy feature of the Cascading Glacier is that on reaching the end of its high valley, at an elevation of 1500 feet or more above the fiord, its slope abruptly changes and it descends the steep south wall of Nunatak Fiord in a series of steps, with greatly-broken surface (Pl. LX, B), like a frozen waterfall on the mountain side, which, from its resemblance to a cascade, led Dr. Gilbert to give this glacier its name. The slope over which this glacier descends is so steep that it is doubtful if the glacier could cling to it were it not for the fact that there are two pronounced rock terraces and several smaller ones (Pl. LXXII, A), caused by the fluting of the valley side by earlier glacial erosion, when the former expanded Nunatak Glacier rose up over the valley side

[1] Gilbert, G. K., Harriman Alaska Expedition, Vol. 3, 1904, p. 60.

[2] Gannett, Henry, Same, Pl. III opposite p. 58.

and swept powerfully westward through the fiord. The ice of Cascading Glacier pours over the lip of its valley down to a point just below the second pronounced rock bench (Pl. LXIII), where its further advance is checked, partly by ablation and partly by the discharge of fragments that tumble down from its terminus. The melting of the glacier has produced no single large stream, but a multitude of small rills extend from its front down to the cliff base. These tiny streams have not sufficient volume, and have not been at work for a long enough time to have erased the glacial grooves produced by the formerly-expanded, westward-moving, Nunatak Glacier. The largest stream, itself of small size, emerges from the west end where the ice rests on the upper rock bench. Apparently the downward extension of Cascade Glacier is mainly checked by the falling of ice blocks rather than by ablation. During each of our visits to this glacier we either heard or saw the fall of the ice blocks at frequent intervals. Usually these falls consist of only small masses, but oftentimes great blocks were detached. There is an extensive talus of these fallen ice blocks, mixed with rock fragments, rising almost up to the terminus of the glacier.

By these falls of ice the form of the lower edge of the Cascading Glacier has been materially altered since it was first observed by Gilbert in 1899, the greatest part of the change having occurred between 1899 and 1905, though there have been noticeable changes between 1905 and 1909. In 1899 a broad tongue of ice extended from the lower bench down the steep cliff face to a moraine area at its base, which was probably the lateral moraine of the land tongue of Nunatak Glacier. This projecting point of Cascading Glacier had almost entirely disappeared in 1909 and the total recession at this point has amounted to between 200 and 300 feet. Its form changed slightly between 1909 and 1913. There have been minor changes in other parts of the Cascading Glacier margin, but nothing noteworthy excepting the developing of a moraine-covered area on the west side just above the lower bench.

There is no sign of response of this glacier to advance through earthquake shaking, and in view of the fact that it has been examined and photographed in 1899, 1905, 1909, 1910 and 1913, each of the latter four years showing recession, we feel warranted in assuming that this glacier has not made any advance whose effect extended beyond the lip of the mountain valley. In view of the shortness of the glacier we assume that it will not in the future respond to the influence of earthquake shaking. Why this glacier should not respond, as other glaciers have done, is not easy to understand, unless it is due to the fact that in this high-lying valley, in the midst of mountains which do not rise to any great height above it, there are not extensive slopes on which the snow rests with sufficient instability to be shaken into the valley in great enough quantities to cause an advance in the glacier.

Significance of the Cascading Condition. Cascading Glacier is one of many glaciers of the same type, found not only in the Yakutat Bay region, but in many other parts of the glacial region of Alaska. Indeed, the name has suggested itself to others besides Gilbert, and there are already at least three other glaciers of this name in Alaska, beside glaciers with names like Cataract Glacier, Hanging Glacier and Toboggan Glacier, which are evidently of the same type. This cascading condition seems to be dependent upon two factors widespread in Alaska; first, the former extension of great trunk glaciers which have so lowered the main valleys by glacial erosion as to leave the tributaries hanging high above their bottoms; and, secondly, the uncovering of the steepened slopes below the

lips of the hanging valleys by the recession of the trunk glaciers. Applying this explanation specifically to the Cascading Glacier of Nunatak Fiord, the evidence is conclusive that the entire fiord, up to a level much higher than the point where Cascading Glacier emerges from its valley, was formerly filled with ice which moved vigorously westward. The deposits left by this expanded Nunatak Glacier, and the grooves and flutings which it wore in the rock of the mountain sides are plainly visible throughout the fiord. That this greatly-expanded glacier flowed vigorously, and for a long period of time, is proved by the series of perfect hanging valleys perched high above the water surface on both sides of the main fiord. The inference is, therefore, warranted that before Nunatak Glacier reached this expanded condition there was a main valley whose bottom was at least as high as the bottoms of the hanging valleys, and that while the lateral valleys have been worn by glacial erosion to some extent, the main valley has been worn much faster and much deeper.

During the ice-flood condition of the Nunatak Glacier these lateral glaciers presumably entered the main glacier with their surface levels approximately accordant with the surface of the main glacier. Such a condition is now observed in tributaries of the larger glaciers far back in the mountains. In these cases, however, there is usually a step or descent in the glacier surface just above where the tributary joins the main glacier. Whether this condition is due to a moderate recession of the main glacier, or whether it is normal to the flooded stage, cannot be stated on the basis of direct observation, but by inference it seems probable that in the stage of highest ice flood the tributary glacier surface would be approximately accordant with that of the main glacier and that the step would be absent. During the flood stage the tributary ice streams contribute a greater or less amount of ice to the main glacier according to their size; but in the great majority of cases these tributaries are so small that as soon as they join the main stream they are at once dominated by the powerful thrust of the main glacier. That this was true of the Cascading Glacier is clearly proved by the huge parallel rock grooves on the mountain slope beneath it. These were without question eroded by the vigorous westward-moving Nunatak Glacier, and there is no sign of grooving resulting from the incoming of the Cascading Glacier.

When the main glacier begins to wane, its thickness decreases while, at the same time, the front recedes up-stream, and thus ultimately, because of thinning, the fact of discordance between the bottom levels of the tributary and main valleys become apparent by the development of a step where the tributary emerges to join the main ice stream. With continued lowering of the main glacier the length of this step increases, and we may ultimately have a cascading glacier extending from the lip of its valley down almost to the base of the steepened slope beneath the hanging valley. This stage in the development of cascading glaciers is illustrated by a tributary on the south side of the Nunatak Glacier about a mile east of the Cascading Glacier. It descends from a valley of about the same height as that of the Cascading Glacier, but its cascading front extends clear down to the glacier surface. Scores of instances of similar condition are found in the Alaskan region. The next step consists in the discontinuation of the tributary end, and then we have the production of the typical cascading glacier, of which the one under consideration is a perfect example.

Once the cascading glacier is disconnected from the main ice stream to which it was formerly tributary, its recession is relatively rapid. This is due first to the steepness

of the cascading part of the glacier, which permits the falling of ice fragments from its front, thus adding a very effective cause for recession to the ordinary one of melting. A second important reason for more rapid recession in this stage is oftentimes the withdrawal of the main glacier from beneath the cascading glacier. This withdrawal of the ice results in the raising of the mean annual temperature in the immediate neighborhood, and consequently in an increase in ablation.[1] This cause becomes particularly effective when by the withdrawal of the main glacier the waters of a fiord take its place. Perhaps this is the main reason why the Hanging Glacier on the north side of Nunatak Fiord no longer cascades out of its mountain valley. For these two reasons, after disconnection of a cascading glacier from the main glacier there is a period of rapid recession; but above a certain level melting must again become retarded and recession diminish because of the lower mean temperature at that elevation. Ultimately, if no change in the ice supply comes about, the time might arrive when a balance between ablation and supply would be reached and the end of the cascading glacier hold its position somewhere below the level of the valley lip. Such condition does not seem to be possible in the Yakutat Bay region excepting up those main valleys which are still occupied by ice. Throughout the region, wherever the main glacier has receded past the mouths of the hanging valleys, it is apparent that but a short time has been required for recession to destroy the cascading condition and to force the ice front back up into the hanging mountain valley. This fact is also illustrated in the Nunatak Fiord. Almost directly opposite the Cascading Glacier, on the north side of the Nunatak Fiord valley, there is a hanging valley of somewhat lower level than that occupied by the Cascading Glacier, and, therefore, presumably formerly occupied by a more vigorous glacier. The first positive knowledge in regard to this valley was obtained from Gilbert's photograph in 1899 (Pl. LXIV) where it is seen that the tidal arm of Nunatak Glacier at that time expanded completely across the mouth of this hanging valley. In the valley was Hanging Glacier, an ice tongue in which clear ice showed just above the lip of the hanging valley, while below the lip a moraine-covered tongue projected to about half the distance between the lip and the surface of Nunatak Glacier. Because of the lowness of this hanging valley, and because of the fact that it opens toward the south, ablation had proceeded much faster on the end of this glacier than on the Cascading Glacier. It was, therefore, so covered by moraine that the appearance of cascading was entirely absent, a condition due partly to the fact that the slope below the lip of the hanging valley is less steep than that in the Cascading Glacier, partly to the presence of an extensive deposit of gravels on the northern side of Nunatak valley rising part way up the mountain base. In the interval between 1899 and 1909 Hanging Glacier receded so perceptibly that its lower moraine-covered end was almost up to the level of the hanging valley lip; and in the meantime the clear ice area was entirely replaced by moraine-covered ice. At the present rate of recession it cannot be long before this glacier will lie entirely inside the lip of the hanging valley. It may, therefore, be considered a third stage in the destruction of the cascading type of glacier.

Succeeding stages introduce no points of special significance, for the later stages consist merely in continued recession of a valley glacier. In the valley next west of the Cascading Glacier, which hangs about 700 feet above the fiord level (Pl. LI, B), there is such a receding glacier whose end now lies about a mile back from the lip of the hanging valley. In this case a single good-sized glacial stream flows from the ice front down to

[1]The covering of a main glacier by a thick mantle of ablation moraine during recession has the same effect.

Plate XLIX

Southeastern Margin of Hubbard Glacier in 1909

Plate L

PLATE LI

A. INTERIOR ALLUVIAL FLAT OF VARIEGATED GLACIER

Ice underlies the crescentic moraine on outer side of flat. Alluvial fan building stream emerges from flat through gorge in the moraine-covered ice in left-hand third of view. Site of this photograph (taken August 8, 1905) was, in 1906, a bristling, crevassed ice front.

B. THE BUTLER GLACIER BULB IN 1909, IN FOREGROUND

In distance Nunatak Fiord, Nunatak and Cascading Glaciers, and hanging valley to the right of Mt. Draper.

VARIEGATED GLACIER IN 1909

Moraine-covered inner bulb of Variegated Glacier in 1909 from Station H (Map 3). Compare with Plate LVII, A, and note disappearance of gorge and waterfall through advance in 1906, and change in glacial streams in interior flat.

VARIEGATED GLACIER IN 1905

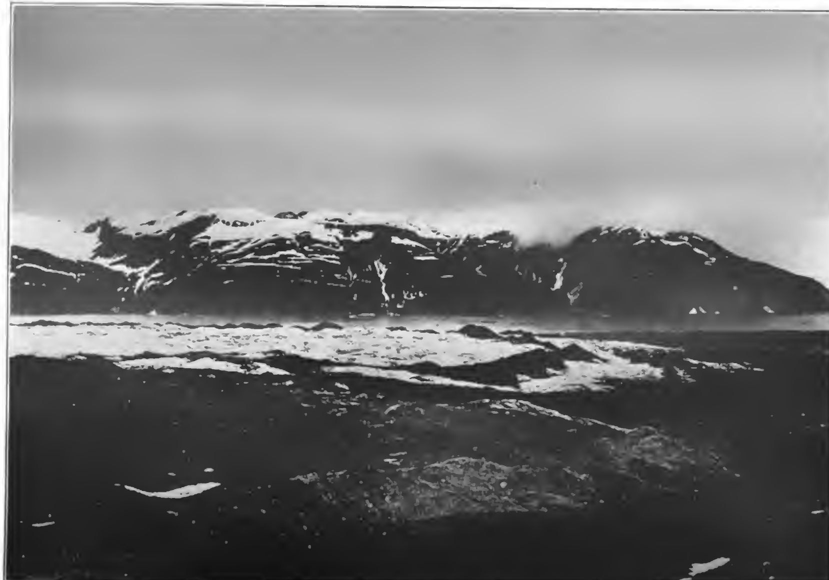
PLATE LIV

SURFACE OF VARIEGATED GLACIER IN 1909
From a point near the site of Plate LIII. Note increase in thickness of glacier.

PLATE LV

SURFACE OF VARIEGATED GLACIER IN 1909

PLATE LVI

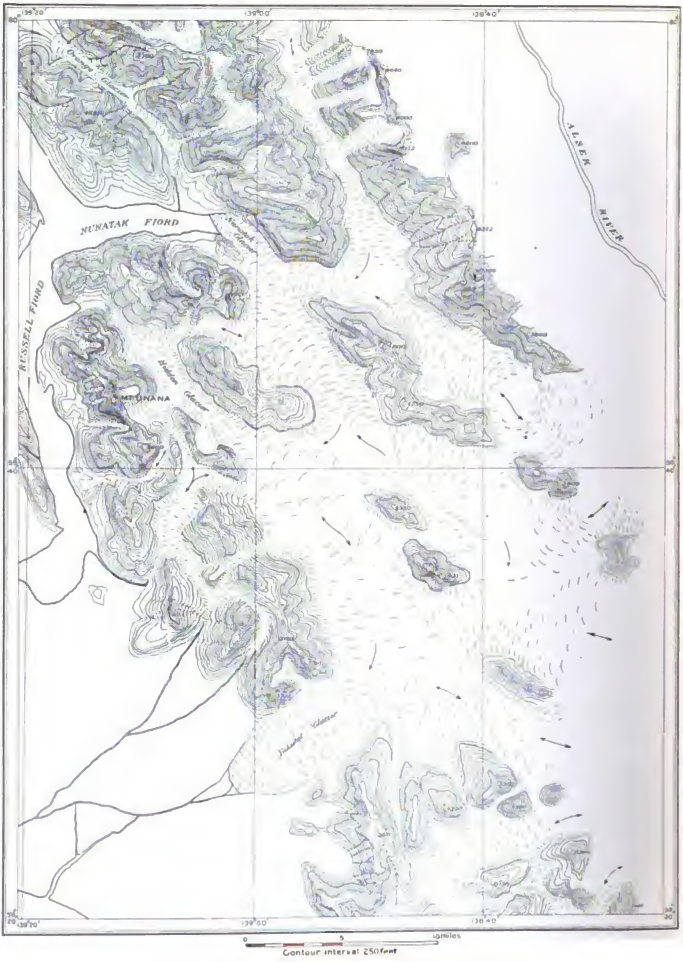


CLEAR ICE IN BULB OF VARIEGATED GLACIER IN 1909

A. Variegated Glacier in 1905

From Station II (Map 3). The advance in 1906 covered gorge and waterfall. See Plate LII and ink line near date 1909.

B. Gap Between the Outer Moraine-Covered Ice and the Butler Glacier in 1909



Nunatak and Hidden Glaciers and Their Snowfields
Map by Canadian Boundary Commission.

PLATE LIX

FRONT OF NUNATAK GLACIER IN 1906 FROM CREST OF NUNATAK

A. Medial Moraines on Nunatak Glacier in 1905
From crest of nunatak. A through glacier on right.

A. Nunatak Glacier and its Land Arm
Photograph, 1906, by Canadian Boundary Survey.

B. The Land Tongue of Nunatak Glacier
From Station C (Map 4), showing recession from 1899 to 1909

A. LAND ARM OF NUNATAK GLACIER IN 1909

B. PROSPECTORS' 1898 SLEDS, ETC., NEAR END OF NUNATAK GLACIER IN 1909

Plate LXIII

Cascading Glacier
Photograph, June 21, 1899, from Station B (Map 4), by G. K. Gilbert.

TIDAL ARM OF NUNATAK GLACIER IN 1899
Photograph by G. K. Gilbert from Station A (Map 4). See also three views from same site in later years (Plate LXV).

the edge of the hanging valley (Pl. LXXI), then tumbles precipitously down the slope in a shallow gorge (Pl. LX, B), which it is actively deepening. In the earlier stages of destruction of a cascading glacier the drainage is less concentrated, and when we reach as early a stage as that of the Cascading Glacier itself, a multitude of streams discharge the ice drainage down the steepened slope in no well defined courses.

This cascading type of glacier, exists, to some extent at least, in the Alps, and, in fact, must be a common condition in the recession of trunk glaciers. Doubtless it was far more common, and far better developed in the Alps during the greatly expanded condition of glaciers in the Glacial Period; for the Alpine glaciation of that time more closely resembled the glaciation in Alaska to-day than does the system of present-day Alpine glaciers.

CHAPTER IX

THE HIDDEN, FOURTH, AND SMALLER GLACIERS

The Hidden Glacier

General Description. The Hidden Glacier was named by Russell [1] in 1891 when he was exploring Russell Fiord and "sailed slowly southward before an uncertain breeze, and about five miles south of Cape Enchantment saw a deep opening in the steep bluffs forming the eastern wall of the valley. A curve in the shore there forms a shallow bay, at the head of which there is a break in the hills, and we could look into the mouth of the canon-like valley which comes down to the water with a very low grade, and is occupied a short distance within by the end of a good-sized glacier. Only glimpses of this interesting valley and of the glacier which it shelters, named Hidden Glacier, could be had as we passed."

This glacier was photographed and roughly mapped by the Canadian Boundary Survey in 1895 [2] and in 1899 Gilbert and Gannett, of the Harriman Expedition, studied and mapped Hidden Glacier,[3] which was revisited and studied by us in 1905, 1906, 1909, 1910, and 1913.[4] No doubt the name Hidden suggested itself to Russell because the glacier was not visible as he sailed up the fiord, excepting from one point of view. Down to 1906 Hidden Glacier retained the same general position, its front being separated from the fiord by an outwash gravel plain a little over two miles in length, terminating in a branching, muddy delta front at the head of a small bay called Seal Bay by the natives; but between 1906 and 1909 the glacier changed greatly. For comparison of Hidden Glacier as subsequently observed by us, as well as for comparison with its condition when studied by Gilbert in 1899, we will describe its appearance in 1905 in some detail.

The upper parts of Hidden Glacier are better known than any of the other large glaciers of Yakutat Bay. The maps of the Canadian and American boundary surveyors (Fig. 12) show it to be a through glacier connected toward the east with one or more glaciers tributary to the Alsek valley, on the north with the upper portions of the Nunatak Glacier, and on the south with the upper portions of the Fourth and Yakutat Glaciers. We do not know from direct observation what the condition of the glacier tributaries may be, nor their size or number; but the inactivity of Hidden Glacier in 1899, 1905, and 1906, and the condition and position of the glacier front suggests that there is a limited supply of ice. In 1905 the glacier was about a mile in width at its

[1] Russell, I. C., Second Expedition to Mt. St. Elias, 13th Ann. Rept., U. S. Geol. Survey, 1892, p. 87.

[2] Atlas of Award, Alaskan Boundary Tribunal, Sheet 21.

[3] Harriman Alaska Expedition, Vol. III, 1904, pp. 53–58 and Pl. IV.

[4] Tarr, R. S. and Martin, Lawrence, Glaciers and Glaciation of Yakutat Bay, Alaska, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 150, 151; Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 59–64; Tarr, R. S. and Martin, Lawrence, The National Geographic Society's Alaskan Expedition of 1909, Nat. Geog. Mag., Vol. XXI, 1910, pp. 26–28, 42, 43, 45.

terminus, being, therefore, about as wide as Nunatak Glacier or Turner Glacier within its mountain valley. Its surface was not notably crevassed and its front did not extend to the sea. Snow-capped mountains rise above it, on either side (Pl. LXXIII), and a number of small tributary glaciers descend from extensive snowfields on the slopes of the enclosing mountains; but there is no proof of the existence of any large tributaries.

In 1905 the lower five or six miles of Hidden Glacier reminded one very much of the Orange and Fourth Glaciers. The glacier surface was smooth and only slightly crevassed, and there was an almost total absence of morainic débris. On each side of the lower glacier there was a well-defined lateral moraine which in its lower portion rose in the form of an irregular ridge because of the protection which the moraine cover gave to the underlying ice. There were two medial moraines, one coming down to the front close by the southern lateral moraine, the other a few hundred yards farther north. These two medial moraines are interpreted as representing the incoming of two weaker tributaries from the south side and further as indication that the main glacier supply is either from the east or from the north, presumably the former because of the narrow belt of mountains on the north side between the Nunatak and Hidden Glaciers. The lateral moraines extended a short distance beyond the visible front of the Hidden Glacier, forming a hummocky group of ice-cored morainic hills. Between the lateral and medial moraines, even at the very front of the glacier, the surface was remarkably free from moraine, only a small amount of staining in the outermost portion discoloring the otherwise clear ice surface.

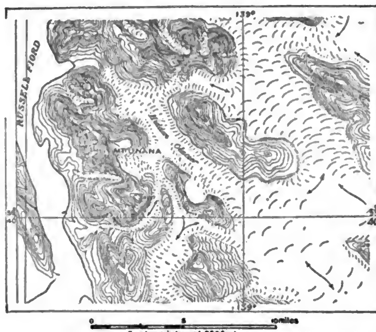


FIG. 12. THE WHOLE OF HIDDEN GLACIER. FRONT SHOWN AS MAPPED BY CANADIAN BOUNDARY SURVEY IN 1895.

The lowest tributary to the Hidden Glacier entered it about half a mile from its front. This tributary is short and is supplied from a great snowfield that clothes the steep upper portion of its valley; but it was evidently not supplying a great amount of ice to Hidden Glacier in 1905 for its lower portion was completely covered by ablation moraine. Ablation was in rapid progress all over the lower part of the Hidden Glacier, and the snow line lay far up the glacier.

The surface slope of Hidden Glacier was very moderate, but at the front the slope increased to from 10° to 20°. Here the glacier *apparently* terminated and from the visible front two fair-sized glacial streams emerged, besides several smaller ones. The smaller of the two streams came from the ice on the north side of the valley, entered a deep gorge cut in older glacial gravels, then flowed out in numerous branches over an alluvial fan which graded down to the outwash gravel plain that lay between the front of Hidden Glacier and the sea. According to the Boundary Commission map of 1895 (Fig. 12), and the map made by Gannett in 1899 (Pl. LXXXV, A), the stream on the north side

was the largest, but in 1905 and 1906 the stream issuing from the south margin of the glacier was many times larger than that from the northern margin.

A possible explanation of this condition is suggested by a photograph from the north side of the valley made by Gilbert in 1899. In this photograph the surface of the outwash gravel plain in front of the glacier is seen to rise perceptibly toward the south valley wall; and above a part of its surface, about a quarter of a mile from the glacier front a black area of irregular form rose above the plain (Pl. LXXII, B). This black elevated area was evidently lateral moraine, a continuation of the lateral moraine on the glacier, rising above the outwash gravel plain. Neither the black moraine nor the rising outwash gravel was present in 1905. We assume, therefore, that there was ice here in 1899 which before 1905 had so melted away as to allow the burial of the moraine, and the development of a large glacial stream on the south side.

The south stream emerged in 1905 from a large tunnel beneath an ice precipice (Pl. LXXIV, A), boiling out in great volume and flowing with great velocity (Pl. LXXIV, B) to the sea. It was so clouded with sediment as to be yellowish-brown in color, and was carrying along with it good-sized pebbles and small bowlders. Excepting near the sea, this stream did not possess any distributaries of large size during the period of our observation, differing in this respect from the northern stream, which divided and subdivided into many branches. However, that the south stream was not confined to a single channel throughout the entire season was made clear by the large numbers of branching stream courses on the outwash gravel plain, and also by the entire absence of vegetation on this plain. It was evident that during the spring melting the volume of both the south and north streams was so increased that they sent branches in constantly-shifting courses over all parts of the outwash gravel plain.

The Outwash Gravel Plain. It is these streams that have built the extensive gravel plain that lay between the glacier and Seal Bay in 1899 (Pl. LXXV, A), 1905, and 1906. In those years the streams were actively engaged in building both upward and outward, as they doubtless had been for many years before. On the seaward margin the plain stood at and below sea level, a broad, muddy, tidal flat being exposed at low tide, and shallow water extending some distance from the visible front of the delta; but at a distance of a few hundred feet from low tide line the submerged edge of the delta was reached, the front sloping abruptly to the deep water that occupied outer Seal Bay. From its tidal margin to its inner edge, at the glacier front, the surface of the outwash gravel plain, or valley train, was for the most part smooth, excepting where crossed by occupied or abandoned stream channels. The slope up to the glacier front was almost imperceptible and the elevation of the inner portion of the plain was only about 150 feet.

The inner portion of the outwash gravel plain, close by the ice front, presented peculiar and interesting phenomena. Fringing the glacier front was a broad ditch, or fosse, (Fig. 13) about 20 yards in width, terminated on each end by a series of low, hummocky ice knolls with gravel veneer, having the general appearance of groups of kame hills, with crescentic form paralleling the glacier front. One wall of this fosse was formed by the Hidden Glacier front (Pl. LXXV, B), which rose at an angle of about 15°. The other wall, which was about 20 feet high, was a gravel-veneered ice cliff with an angle of slope greater than 30°. The presence of ice in this wall of the fosse was shown in a number of places where water escaped through small caves in the ice cliff. Ice also

existed on the floor of the fosse, thus proving that Hidden Glacier, apparently ending in a clean ice slope at an angle of 15° or 20°, in reality extended out beneath the fosse and under the gravel plain. The presence of ice even farther out was plainly shown by the pitted condition of the outwash gravel plain surface, as was observed by Gilbert [1] in 1899 before the fosse had developed; and it was exhibited with remarkable perfection in 1905. Some of the kettles were small, shallow pits a few feet in diameter. Others were of large size, the largest measuring 250 feet in length and 50 feet in width. The depth of the kettles varied from 1 or 2 feet to 15 or 18 feet. Many of the kettles, especially the shallow ones, were dry; but others, notably the large kettles, contained pools (Pl. LXXVI) of clear, deep-blue water a few feet in depth, the greatest depth measured being about 10 feet. Altogether there were over a hundred good-sized pits in

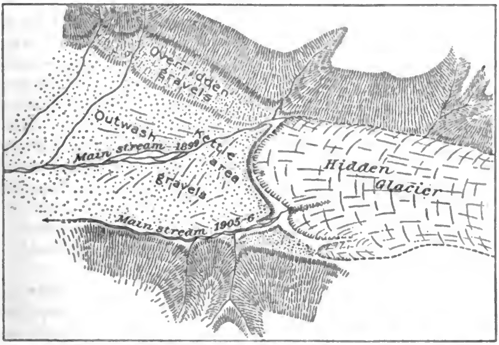


FIG. 13. HIDDEN GLACIER IN 1905, SHOWING FOSSE BETWEEN THE GLACIER AND THE WORDS *Kettle area.*

an area of about a square mile, and the region pitted by these kettles extended fully a mile from the visible glacier front.

The presence of the kettles in the outwash gravel plain was interpreted by Gilbert, and by ourselves, as proof of the presence of ice beneath the gravels. There is abundant evidence to support this conclusion. In the first place, the great amount of clear, ice-cold water emerging from the kettles could be accounted for only by the melting out of buried ice. Numerous small streams issued from the kettles, their clear water contrasting strikingly with the muddy water of the nearby glacial streams. That the kettles were developing by irregular subsidence of the surface of the outwash gravel plain during our visit in 1905 was clearly proved by the abundant evidence of faulting and slumping of the gravels and the sliding down of the kettle walls. Furthermore, on the bottom of some of the kettles were deposits of mud recently made, and faulted by subsidence since their accumulation. The very existence of the kettles on the gravel plain is indication of the recency of their origin, for the surface of the plain was traversed by an intricate series of abandoned channelways, some of the largest with banks 30 or

[1] Gilbert, G. K., Glaciers and Glaciation, Harriman Alaska Expedition, Vol. 3, 1904, p. 64.

40 feet apart, and with a depth of 5 to 10 feet. There is every reason for believing that these channelways had been occupied by water derived from the melting of the snows in the spring of 1905. Further proof that the surface of almost all parts of the gravel plain was frequently reached by the branching glacial streams is found in the absence of vegetation, for even individual annual plants were found in only a few places. With such rapid upbuilding of the plain as was evidently in progress, and with such recent extension of the streams over its surface, it is inconceivable that kettles could long remain, for they would soon be destroyed by stream erosion and by gravel deposition. We are convinced, therefore, that the kettles observed in 1905 were in the main the result of the melting of the buried ice in that single season.

The presence of buried ice beneath the outwash gravel plain, and its direct connection with Hidden Glacier, are established by the facts observed in 1905. We could not determine exactly how far the ice extended, though the presence of kettles to a distance of at least a mile from the ice front furnished evidence of its extension that far. Beyond this point no kettles were observed, and while it is possible that ice may have existed further than a mile from the visible ice front, it is not probable, and, if it did, it was certainly very thin. The inference that we draw from the facts observed, is that a moderately-thick ice foot, buried beneath the gravels, projected for approximately a mile beyond the visible front of Hidden Glacier, gradually thinning and disappearing. There was a continuous diminution in the size of kettles and in the degree of slumping of the surface for about a mile from the ice front. This of course may in part have been due to deeper burial of the ice on the outer portion, but it was probably mainly due to the thinning of the ice. It is a noteworthy fact that the area of most abundant and largest kettles, which was within a quarter of a mile of the visible glacier front, was on the site occupied by the glacier when Gilbert photographed it in 1899.

The presence of this buried ice beneath the gravel plain, and the evidence of its direct connection with the glacier beneath the fosse, supports the conclusion reached above that Hidden Glacier was not in active movement in 1905, at least in its outermost portion. It would have required only a very slight thrust to have broken the buried ice, and with it the overlying gravel plain. Even at the very edge of the fosse the gravel showed no sign whatsoever of faulting, with the exception of that due to subsidence. We infer from this fact, as well as from the uncrevassed glacier surface and the evidence of recession of its visible front, that in 1905 the lower part of Hidden Glacier was in a stagnant state, and that its recession would in all probability have continued, as in the past years, were it not for the absolute change in conditions brought about by the earthquakes in 1899.

The manner in which we conceive the burial of the glacier terminus by the gravels to have been effected is as follows: In the early spring the expanded surface of Hidden Glacier reached at least as far out as the west wall of the fosse, and probably farther. With the coming of spring, large streams, derived from melting of snow upon the glacier deposited gravel upon the glacier terminus, thus protecting it from melting rapidly. When such streams waned and ceased depositing, ablation would proceed more rapidly upon the clear ice portion of the glacier than upon that veneered with gravel. In this way a fosse or valley developed between the clear and the gravel-veneered ice. A lake might form in such a depression, ultimately being drained through moulins. Where the gravel was thinnest, that is nearest the glacier, the subsequent, more rapid melting of the ice beneath the gravels resulted in slumping, with the production of a kame moraine.

Farther out only pits were formed at first, being filled later by alluviation; and this process might go on for years. The ultimate kettles upon the pitted plain would be those due to the melting of the last isolated remnants of the buried glacier that were not filled by the last alluviation.

The outwash gravel plain built in front of the Hidden Glacier furnishes an excellent example of a deposit common in regions of Pleistocene glaciation in Europe and America, but here in actual process of formation. The significance of the phenomena observed in this valley in 1905 will be apparent to students of Pleistocene glacial deposits. If the ice had all melted away in 1905 there would have developed a crescentic kame moraine area with an ice-contact face toward the glacier and with a sloping outwash gravel plain toward the sea, pitted by numerous kettles. The stream channels on the plain would extend up to where the glacier front stood in 1905. Many of the kettles would be filled.

Recession Between 1899 and 1905. The photographs and map made by the Harriman Expedition in 1899 give a basis for fairly exact comparisons of the conditions in 1899 and 1905. In some cases it was possible to occupy the exact sites of Gilbert's 1899 pictures (Pl. LXXVII); in other cases the approximate position could be located. In the interval of six years there was a recession of the Hidden Glacier front of more than a quarter of a mile. There was evidently at the same time a change in position of the most projecting part of the visible ice front, for Gannett's map shows the outermost point close to the south wall of the valley, whereas in 1905 the glacier projected most near its center. There was also a change in the point of emergence of the largest glacial stream, as already stated. The lateral moraine on the south side was more pronounced in 1905 than in 1899, but this fact is probably only partly due to continued recession, for our visit was about a month later in the season than Gilbert's, and naturally by that time summer ablation would have brought the moraine into greater relief.

Earlier Recession. Prior to 1899 we have no sufficiently accurate basis for determining the position of the Hidden Glacier front, though the Canadian Boundary Commission map, based upon observations in 1895, and Russell's name Hidden Glacier, given in 1891, both indicate that the glacier front was in those years well back within the mountain valley. The glacier was less than 1¾ miles from the fiord on the 1895 map, and, if this is accurate, it retreated about 500 feet between 1895 and 1899. However, that Hidden Glacier had recently been much more extensive than in 1899 was clearly proved by a number of facts, notably the pronounced grooving and fluting of the mountain slopes, to an elevation of 1000 or 1500 feet, of such recent date that weathering had not notably erased it; the general absence of vegetation on the lower valley slopes and the entire absence of alder thickets; and the presence of overridden gravels.

The north side of Hidden Glacier valley, which has not been as greatly oversteepened by glacial erosion as the south side, had in 1899 and 1905 a series of gravel deposits extending two or three hundred feet above the outwash plain and down the valley for a mile and a half. The gravel surface was sculptured in indistinct, westward-descending terraces, and stream cuts showed excellent stratification with fragments of trees among the gravels.

The north end of the glacier rested upon these gravels (Pl. LXXX, B), the glacier surface sloping 10° to 15° westward and the gravel surface about 5° southeastward, or toward the centre of the glacier, causing the ice to wedge out into a thin edge. It was

thus visibly demonstrable that the glacier had overridden the gravels, eroding them, as shown by the broadly-truncated stratification of the gravel, and with erosion most pronounced in the centre of the valley because the eroded surface sloped that way. Gilbert had detected the overriding of these gravels in 1899,[1] before recession had revealed the site where the glacier itself rested on the gravels in 1905 (Pl. LXXVIII, A), observing the smooth curves eroded by flowing ice, truncating stratified gravel deposits. He also noted the fact, which we likewise observed, that there was a veneer of till and scattered erratic blocks resting on the surface of the truncated gravels. The overridden gravels extend outside of Hidden Glacier valley far down into Russell Fiord, proving a former great extension of the glacier, during whose advance erosion had sculptured but not removed gravels of earlier deposit.

In 1905 a stagnant, outlying mass of moraine-protected ice lay upon the gravels over an eighth of a mile from the glacier, showing gradations of kame topography, with various stages of melting according to the distance from the main glacier. It was a lateral moraine mass whose melting would later produce a superposition of hummocky topography upon earlier, ice-eroded gravels.

Thus in 1905 the terminus of Hidden Glacier showed clearly the two interesting phenomena of a glacier margin resting upon old gravels which were sculptured but not wholly removed, and the middle of a glacier with its terminus buried beneath modern gravels which were still being deposited upon it.

Condition in 1906. The recession of Hidden Glacier from its former advanced position, concerning which we have specific evidence between 1899 and 1905, continued at least until June, 1906. At that time the Geological Survey party visited Seal Bay and took a photograph of the Hidden Glacier from near its mouth, but seeing that the glacier was in essentially the same position and condition as in 1905, and having other objects in view for the season's work, no further study was given Hidden Glacier at that time.

Condition in 1909. On July 10, 1909, when the National Geographic Society party rounded the point which forms the north side of Seal Bay, the formerly Hidden Glacier burst upon our view with an ice front so high and so near that at first we thought it had become tidal (Pl. LXXVIII, B). It no longer deserved the name Hidden, for it boldly projected out in its valley so far that it formed a prominent feature in the landscape from all points of view where one could look into Seal Bay. Hidden Glacier, stagnant in 1899, 1905, and as late as July 6, 1906, and showing striking evidence of notable recession between 1899 and 1905, had suddenly changed its condition, and between July 6, 1906, and July 10, 1909, had pushed its front forward fully two miles, and to within a little over a quarter of a mile of the sea. Where the ice front stood in 1905 and 1906 there was a thickness of about eleven hundred feet of ice in 1909 (Fig. 14), as shown on our detailed topographic map (Map 5). Our most important photographic sites, selected in 1905 for the purpose of showing changes in the glacier front as it continued to recede, were deeply buried beneath the glacier. The formerly-overridden gravel terraces were covered by ice and the glacier had completely overridden the pitted outwash gravel plain (Pl. LXXXV). Its front lay nearly a mile beyond the outermost limit of kettle development in 1905.

Two of our photographic sites of previous years could be reoccupied because they lay far enough down the valley to be beyond the reach of the advancing ice front. One of

[1] Gilbert, G. K., Glaciers and Glaciation, Harriman Alaska Expedition, Vol. 3, 1904, p. 58.

these sites was on a low knoll, about 200 feet above the sea, and very near the outer edge of the outwash gravel plain. In the picture of July 26, 1905, from this site the glacier front is distant and ice occupies only a very small fraction of the entire picture, all the foreground for two miles being outwash gravel plain. In the picture of July 10, 1909, from the same site, on the other hand, the glacier dominates the view (Pl. LXXIX, A), and the outwash gravel plain occupies only a narrow strip. Between two and three square miles of ice, with an average depth of several hundred feet, has been added in this picture. No less striking is the difference in appearance of the Hidden Glacier valley in the two photographs from the same site on the alluvial fan at the outermost part of the outwash gravel plain, one taken July 6, 1906 (Pl. LXXX, A), and the other July 10, 1909. In the latter picture the ice front is near at hand and the glacier is so large that it was impossible to show it on a single 5 x 7 plate, whereas in the 1906 photograph the glacier front extends only about a fourth of the way across the picture. Such a sudden and enormous change in the position of the glacier front in so brief a time is, so far as we know, without recorded parallel. Accustomed as we were by this time to such transformations of glaciers, the change in the Hidden Glacier seemed to us almost incredible, even as we looked upon it with our own earlier photographs in our hands.

The surface of the glacier contrasted as strikingly with its 1905 condition as did the position of its front. From the very outer edge, back as far as we could see up the glacier valley, and from one side to the other, the ice surface was rough and broken, forming a striking contrast to the smooth, unbroken surface of 1899, 1905, and 1906. The roughness, though great, was not of the same character as that presented by the Atrevida, Variegated, Haenke, and Marvine Glaciers in 1906, and by the Lucia Glacier in 1909. In these cases the ice was so broken, there were so many pinnacles, and such a mass of yawning crevasses, that the glaciers were impassable; but in the Hidden Glacier the irregularities were all rounded (Pl. LXXXI) and it is clear that there had been not only a breaking of the surface, but, as in the case of the Variegated Glacier in its 1909 condition, a subsequent healing as well. Although exceedingly rough, and with numerous crevasses, it seemed to us from the view which we obtained on the north side at an elevation of 600 feet, where we could look over the broken surface, that it would be possible to traverse it in almost any part, though there were areas of excessive crevassing which would undoubtedly make necessary the exercise of great care. We walked over portions of the northern margin and the lower end of the glacier in order to satisfy ourselves as to the exact conditions from near at hand, and what was observed there seemed to be a duplication of the conditions throughout most of the visible glacier surface. These parts of the glacier surface were made up of a series of huge swells, with somewhat sharpened crests, rising from 20 to 50 feet above the broad depressions between them and giving to the surface an undulation strikingly in contrast with the regularity and smoothness in 1905. Everywhere along the margin of the glacier, the ice was badly cracked and seamed, showing the great strain to which it has been subjected; and, in addition, the surface was interrupted by innumerable crevasses, some of them of considerable depth, so that in crossing the glacier it was necessary to follow a winding path in order to avoid the crevasses. Everywhere ablation was in rapid progress and many streams were coursing over the glacier surface, which had almost no morainic veneer to protect it from melting.

As in 1899, 1905, and 1906 the surface of the Hidden Glacier was in 1909 remarkably free of débris so that in all the views one sees mainly clear ice. There was, in fact, even less débris than in previous years, for, with the advance of the glacier, there had been a lateral spreading and a rising on the valley side, as a result of which the lateral moraines had been destroyed. There was, therefore, an absence of the lateral moraines which formed a prominent feature of the glacier margin in previous years. The medial moraine was, however, still present and still over on the south side of the valley; but near the terminus it swung still farther south, and at the very end of the glacier became lateral. The fact that this medial moraine had retained its approximate position furnishes proof that the thrust by which the Hidden Glacier was broken and pushed forward did not come from the glacier tributary on the south by which this medial moraine is supplied. Had there been such a thrust from that glacier this medial moraine would of necessity have been pushed farther out into the glacier than it was in 1905. We assume, therefore, that the thrust came from either the east or the north.

The margins of the glacier were examined with some interest, because the ice had suddenly risen up over gravel and rock slopes, from which ice had for many years been absent. The two margins of the glacier were found to be quite different. Along the northern margin there was a band of barren moraine 200 to 300 feet in width, with no living vegetation but with many fragments of willow, alder and cottonwood. At its upper border, in a number of places, there was a well-defined ridge of bowlders, till, gravel, and plants shoved up by the advance of the glacier (Pl. LXXXIII). It is a perfect illustration of shoved, or push moraine, and in places was faulted and doubly ridged by the thrust. It wound up and down the hillside in irregular lobes, often rising highest on spurs rather than in adjacent gullies. The barren moraine (Pl. LXXXVI, A) between this shoved moraine and the ice shows clearly how much recession there had been since the period of greatest lateral advance. In some places the marginal moraine was distinctly hummocky, especially in places where the hillside drainage, which in earlier years extended down to the outwash gravel plain, found itself interrupted in its progress by the newly-established base level of the advanced glacier. In these places the sediment which the hillside drainage was carrying had in part been deposited on the ice margin, and in some places there were patches of ice, still unmelted, in the barren zone from which elsewhere the glacier had completely receded. All along this margin there was evidence that the ice was rapidly withdrawing, with a thin wedgelike edge resting on the older till and gravels. Not all the barren zone above the present ice margin can be ascribed to recession of the glacier, for some of it doubtless represented the position of lingering snow banks, prevented from sliding down farther into the valley by the newly-imposed ice barrier. But this explanation can account for only a few areas of unusually broad barren zones, for in those areas which lie between the ice and in shoved moraine there can be no question but that ice extended up to the pushed area. Also in those portions of the barren zone where ice still remained beneath gravel deposited by marginal drainage we have positive proof that the barren zone was due to recent occupation by the glacier itself.

On the north side of the glacier the hillside drainage, now interrupted in its journey to the valley bottom by the presence of the glacier, united with the waters supplied by the melting of the glacier and flowed along a marginal course, in places flowing under the glacier for short distances, elsewhere flowing at or near the ice margin. In one place

the marginal stream had cut a distinct rock gorge (Pl. LXXXIV), from 2 to 8 feet in depth, contouring the hillside. It cannot be conceived that on this steep mountain slope a stream would follow the contours for several hundred feet, unless some barrier interfered with its flow down the hillside. Evidently, therefore, this rock gorge was the result of marginal drainage. We cannot be certain that the channel may not have started at some earlier stage in a high level of Hidden Glacier, though the coincidence of position at the exact margin of the present stand of the glacier would in that case be quite remarkable. The shallowness of the gorge, and indications of its newness, have led us to believe that it is the product of marginal drainage at the present stand of Hidden Glacier.

On the south side of the Hidden Glacier valley the mountain walls rise more abruptly, and there is an absence of gravel slopes such as exist on the northern side. Only for that reason the marginal conditions on the south side are quite different from those on the north. There is also the factor of difference in power of sunshine on the two sides of the valley. On the moderately-sloping north side the low-lying sun strikes with power, but on the steeply-rising south side the sun's rays produce little direct effect. Indeed, at the mountain base for a considerable part of the day in summer this side of the valley is in shadow. As a result of these different conditions on the two sides, there was as yet no marginal valley there, and the ice extended with uniform slope right up to the mountain base. In fact, as late as the middle of July, there was a slope upward from the ice surface caused by the snow banks which had slid from the mountain side down to the glacier, which had not yet melted. By comparison of the photographs of 1909 with those of the previous years it became evident that the amount of snow remaining on the mountain slopes just above the glacier was much greater in the latter year than in the earlier years. The advance of the glacier and the rising of its surface introduced a temporary change in the local climate. Both on the south and the north margins there had as yet been too little ablation to have brought out in relief the lateral moraines which will doubtless ultimately develop again on both sides of the Hidden Glacier.

The very front of the Hidden Glacier sloped moderately, almost to the edge, with some low, débris-covered ice cones just back of it on the glacier surface, then ended in a low ice cliff from 20 to 30 feet in height. The ice front projected farthest in the southern half of the valley. From the ice front issued many small streams and two good-sized ones, the larger, as in 1905, issuing from the south side very close to the valley wall. The next largest stream came out 100 yards from the northern margin, and was evidently carrying not only drainage from the ice, but also the marginal drainage of the north side of the glacier which in its lower course flowed through a tunnel in the ice. Fresh faulting by slumping of the ice surface near where this north stream emerged, showed clearly that its drainage was in process of development, and the extension of this slumped area towards the northern margin of the glacier is proof of the direction from which the supply for this stream was derived. Both the south and north streams flowed out upon the gravel plain in front of the glacier, and thence down to the sea; but between these two streams there was a much smaller stream with water derived from a number of small streams issuing from the steep ice front, at first flowing in a shallow valley parallel to the ice front, the streams from the two ends of the valley finally uniting in about the center of the glacier and going to the sea as one torrent.

The southern portion of this frontal valley was an incomplete fosse, presumably of the same origin as that developed in 1905 when the glacier front stood much farther up the valley. The inner boundary of the fosse was the visible glacier front, the outer margin a sloping gravel terrace with a low, undulating slope toward the fosse and a gradual descent of its surface toward Seal Bay. The apex of the terrace was toward the large south stream and was evidently an alluvial fan of that stream, probably built in the spring. Between the southern edge of the fosse and the glacier there was a low morainic mass which was slumping rapidly. This extended northward along almost the entire glacier front, but grew lower and had less pronounced form toward the north. Throughout its entire length this hummocky area had ice beneath it, and the hummocks rose from 5 to 25 feet.

The alluvial fan terrace just described had numerous small, dry kettles, and just in front of the ice, and for some distance northward there were also kettles, some containing water. Ice was revealed in the southern side of the pitted alluvial fan where it was cut into by a swing of the large glacier stream. That it was an alluvial fan, built on the outer edge of the glacier during the period of spring melting was indicated by the presence of numerous channelways upon it, pointing toward the apex of the fan and, therefore, toward the south glacier stream which built it. From the conditions on and in this fan we find complete verification of the explanation previously given of the pitted outwash gravel plain which existed farther up the valley in 1905. In that year the process of burial of the glacier terminus was not so clearly demonstrable as in 1909, but of the 1909 condition there can be no doubt as to the cause. The steep slope of the alluvial fan toward the north and west, and the presence of numerous channelways extending in the same direction, both pointed clearly to the source of the gravel-bearing water; and the discovery of ice beneath the pitted gravels verified by observation the inference previously drawn from the presence of the kettles, that they were the result of subsidence through melting out of buried ice. In the 1909 condition the aggradation by which the ice front was buried beneath the gravels was far less extensive than in 1905, and the area occupied was much smaller, though otherwise the phenomena were essentially the same.

Although not tidal in 1909, the Hidden Glacier discharged ice into Russell Fiord. Several scores of these icebergs were afloat in Seal Bay in July, 1909, some of them three or four feet in diameter. They had floated down the largest glacial stream from the glacier front.

A comparison of maps and photographs of the coast of Seal Bay in 1905 and 1909 shows that the northern side of the outwash plain, which projected most in 1905, had grown almost none in connection with this two mile advance of the glacier, while the south side, which projected most in 1909, had grown forward 900 or 1000 feet since 1905. The southern side is the side where the largest glacial stream entered the bay between 1905 and 1909. There are two photographs showing this coast from exactly the same site in 1905 and 1909, and exhibiting this delta advance graphically, though unfortunately they cannot be compared exactly because we do not know the stages of tide at which they were taken.

Interpretation of the Advance. The condition of Hidden Glacier in 1909, as contrasted with its condition in 1906, proves a very great advance and breaking of the glacier during the interval of three years. In view of the resemblance of the phenomena of the Hidden Glacier to those observed in other glaciers of the region and of the brief interval of time

involved, there can be no doubt but that this advance also was spasmodic and in every important respect similar in character to the advance of the other glaciers, the one notable difference being the much greater advance of the front; for in this respect Hidden Glacier stands unique among all the advancing glaciers. While in all cases there was some advance of the front, and in the Haenke Glacier a notable forward movement, there was nothing to compare with the remarkable change in position of the Hidden Glacier front. Doubtless could the phenomenon of advance have been observed it would have presented only a duplication of the same features shown by the other glaciers; but it would have been most interesting to have observed how rapidly the front of the glacier actually moved forward, and also to have seen what response to the thrust the ice buried beneath the outwash gravel plain made. Upon these questions, however, we cannot now hope to have evidence.

As to the exact time of the advance of the glacier we are also unable to make positive statement. It is known to have come between July 10, 1906, and July 6, 1909. There are numerous indications that the advance came nearer the earlier period than the later, and it may even have come later in the season of 1906 than the period of our visit. We cannot conceive that it occurred later than the summer of 1907, for after the advance and breaking were completed, and before 1909, there had been enough ablation to reduce the crevassed, broken surface of the advancing glacier to a passable condition. The full season of 1908, and the first month or two of 1909, certainly represent none too much time for as much healing of the surface by ablation as has taken place. The surface was rougher in 1909 than that of the Variegated Glacier, whose breaking had been so nearly completed in the summer of 1906 as to give ablation some play in that season while during 1907 and 1908 ablation doubtless dominated in the Variegated Glacier. In view of the greater roughness of the surface of Hidden Glacier we are, therefore, inclined to place its period of advance after the close of the summer season of 1906 and very probably before the period of melting in 1907 had proceeded very far.

In addition to the healing of the surface by ablation, the marginal conditions along the northern side of the glacier are suggestive of the lapse of a period of at least one or two years since the maximum advance. This evidence is partly the notable shrinking of the ice from its most advanced position on the northern slopes, and partly the rock gorge occupied by the marginal channel, The bushes killed by the advance were so brittle in 1909 that limbs snapped readily, suggesting that they had been overturned as early as 1906 or 1907. The condition of the glacier front was also indicative of the lapse of at least one full season since the advance, for, in order to develop the cliffed front of the glacier and the fosse, and in order to lower the frontal slopes sufficiently for the deposit of the alluvial fan described above, would seem to demand at least one complete season of ablation. Altogether, therefore, although we cannot state with definite exactness the time at which the Hidden Glacier advanced, we believe that the evidence from ablation and from frontal and marginal conditions demands that this period should be at least before the beginning of the summer of 1908, and later than the autumn of 1906,—that is, some time in 1907. That the advance was rapid and soon ended is proved by the great change in position and condition, for a very large part of the three years available is required for the observed healing by ablation. In other words, the evidence from Hidden Glacier, though in some respects lacking definiteness, is completely in harmony with that furnished by the other advancing glaciers, about which we have more exact information.

The rate of advance cannot be determined exactly, but the following tentative computation can be made. If it had taken the entire period of a little less than three years, from July 10, 1906, to July 6, 1909, for the ice front to advance the two miles, the rate of forward movement of the ice front, ignoring wasting by melting, would have been nine and seven-tenths feet a day, or faster than Muir Glacier was moving in 1890 when its seven feet of daily movement were compensated for by iceberg discharge from the front. The two mile advance of Hidden Glacier, however, we know to have occurred not in three years, but probably in less than one year and, judging by the known advance of Variegated, Haenke, Atrevida, and Marvine Glaciers, during a very few months. Its rate of forward motion must have been several times the nine and seven-tenths rate, and may even have attained a rate of 30 to 50 feet a day. Although we cannot give the rate of advance exactly, it is easy to see that the advance of Hidden Glacier was far different from normal glacier movements and may be spoken of as a spasmodic rush, or flood.

In connection with the recent advance of Hidden Glacier it is of interest to note that one of its through glacier connections on the Alsek may also have commenced to advance

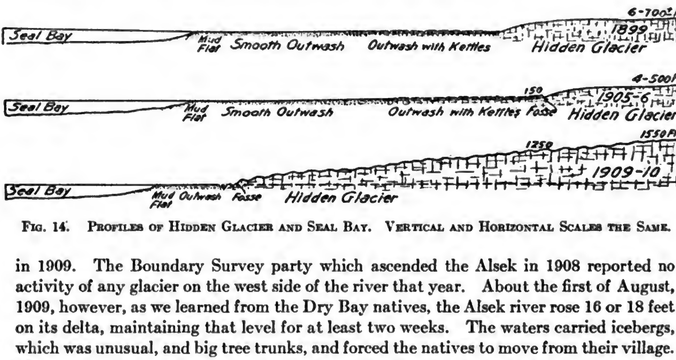


Fig. 14. Profiles of Hidden Glacier and Seal Bay. Vertical and Horizontal Scales the Same.

in 1909. The Boundary Survey party which ascended the Alsek in 1908 reported no activity of any glacier on the west side of the river that year. About the first of August, 1909, however, as we learned from the Dry Bay natives, the Alsek river rose 16 or 18 feet on its delta, maintaining that level for at least two weeks. The waters carried icebergs, which was unusual, and big tree trunks, and forced the natives to move from their village. This rise is interpreted as meaning the crevassing of some glacier up the river, and hence greater melting. The glacier is thought to be one above the lower Alsek canyon, below which no large trees grow, and may well be the large ice tongue on the west bank, next above the lower Alsek Canyon, which connects with the Yakutat, Hidden, and Nunatak Glacier systems. Since equilibrium is disturbed in this glacier system, as evidenced by the two mile advance of Hidden Glacier, it is quite likely that an advance has also occurred on the Alsek, and that other advances may affect the other glaciers of Alsek valley, the Yakutat Glacier and the smaller ice tongues within the next year or two, as it did Nunatak Glacier in 1910.

Conditions in 1910. The junior author found in June, 1910, no particular changes in Hidden Glacier since the year before except the continuation of retreat and thinning by ablation. The drainage had not altered significantly.

Detailed studies of Seal Bay, the submerged continuation of the Hidden Glacier valley,

as revealed by soundings, are considered in Chapter XI. Careful observations showed that the delta advanced 1600 feet between 1899 and 1910, and, as stated on an earlier page, 900 or 1000 feet of this advance came between 1905 and 1909, the increased rate in delta growth during the later period doubtless being due to the large amount of alluvial material supplied during the two mile advance. This rapid rate of growth, 145 to 225 feet a year, is remarkable, for the much larger Mississippi delta advances only 262 feet a year, the Danube 300 to 400 feet, and the Tigris-Euphrates seventy feet annually, though small glacier streams are normally much more heavily loaded with sediment than the rivers mentioned. Seal Bay is 180 to 600 feet deep. This part of Russell Fiord shoreline was probably not uplifted during the 1899 earthquakes, so the 1600 foot advance cannot be ascribed to a change of level of the land. The 1600 foot advance of the Hidden Glacier delta is shown in a spectacular way in Plate LXXXII, where the upper photograph shows the Harriman Expedition vessel, the *George W. Elder*, anchored in June, 1899, at a point shown to be dry land in the lower photograph taken in June, 1910.

Recession 1910 to 1913. In September, 1913, the junior author made more detailed observations of Hidden Glacier than on any other Yakutat Bay glacier visited by the International Geological Congress. The recession of the front of the glacier by melting was estimated to be 400 to 500 feet. The imperfect fosse which existed in 1909 and 1910 was destroyed. The outwash plain in front of the glacier was graded at a lower and flatter level than in 1910, but fragments of the older alluvial fans existed as isolated flattish-topped hills. An abandoned marginal gorge, cut 40 to 60 feet in the older gravels of the north valley wall, no longer had a stream. There were pits in front of the glacier, showing by slumping that buried ice existed on the site of the glacier terminus of 1909–10. Icebergs of small size were floating down the larger glacial streams to the sea. At the northern margin the Hidden Glacier had thinned at least 150 feet vertically and shrunk several hundred yards from the push moraine of the advance of 1907. The crevasses of the glacier surface were largely subdued by ablation. Isolated lateral ice remnants rested on the overridden gravels.

TABULAR STATEMENT OF KNOWN CHANGES IN HIDDEN GLACIER

Year	*Nature and amount of change*	*Recorded by*
Up to 1891	Probably retreat	Russell
1891–1895	Probably continued retreat	Boundary Survey
1895–1899	Retreat about 500 feet	Gilbert
1899–1905	Retreat about 1300 feet	Tarr and Martin
1905–1906	Continued slight retreat	Tarr
1906–1909	Advance over 10,000 feet; slight retreat	Tarr and Martin
1909–1910	Slight retreat	Martin
1910–1913	Retreat, 400–500 feet; thinning 150 feet	Martin

Fourth Glacier

General Description. Until 1909 we had not visited this glacier, though we knew of its existence, partly from the Canadian Boundary Commission map,[1] and partly from visiting the large, milky glacial stream which issues from it and enters the head of Russell Fiord. It was photographed and roughly mapped by the Canadian topographers in 1895 and was well known to the prospectors who crossed it in 1898, and in smaller numbers almost every year since; but they have left us no description of the conditions which they encountered in their journeys over its surface, except the inference that, being passable, it was not greatly crevassed. They went over this glacier highway in sufficient numbers in 1898 to warrant the establishment of a store on the shores of Russell Fiord near the point where the glacial stream emerges, relics of which are still to be seen. Blackwelder saw the glacier in 1906 from a mountain spur some miles south, noting the moraines upon its surface and suggesting a new name (Beasley Glacier) instead of the one generally used. This name[2] has recently been discarded by the U. S. Geographic Board for reasons stated in Chapter II. The glacier was mapped more accurately and photographed from peaks on three sides by the Boundary Survey party in charge of Fremont Morse in 1906, and the sketch map of the glacier (Fig. 15) is reproduced from their unpublished map.

The Fourth Glacier is fed by at least four short tributaries, each about 2½ to 3 miles long, that unite to form the trunk glacier (Pl. LXXXVI, B), which has a length of 3½ miles, a width of over a mile, and terminates approximately 500 feet above sea level, three or four miles east of Russell Fiord. The two Boundary Survey maps show it as a through glacier connected with the upper portions of the Hidden and Yakutat glaciers, which in turn connect with glaciers descending to the Alsek valley. The through glacier divides between it and Hidden Glacier rise 3880 and 3750 feet respectively, that to Yakutat Glacier about 3500 feet. Over this latter pass prospectors state that one could start at the terminus of the Fourth Glacier and find open, ice-filled valleys over which sledging is not difficult, across various divides, and down different glaciers into the Alsek Valley. Fourth Glacier thus has complex relations with other glaciers, because these coast mountains receive such heavy snowfall that the valleys are drowned in ice, forming a network of through glaciers.

Observations in 1909. That part of the Fourth Glacier which lies within the range of our observation is the lower two or three miles, or below where the tributaries from the direction of Hidden Glacier unite with those from the east. In the upper portion of this observed section the glacier emerges from a broad valley enclosed in mountains of no great height, which because of the fact that they face the ocean, and lie not far from it, do not bear a heavy burden of snow. Farther back the mountains rise higher and the snow cover is more extensive. The outer portion of the glacier is made by the union of two pairs of arms, and photographs show three medial moraines that are formed by lateral moraines from these tributaries. Above the terminus of the glacier two medial moraines swing over to the eastern margin and become lateral, the one on the east side being light-colored and that on the left dark-colored. A narrower medial moraine maintains a position near the east side.[3]

[1] Atlas of Award, Alaskan Boundary Tribunal, Sheet 21.

[2] Blackwelder, E., Journ. Geol., Vol. XV, 1907, pp. 417–418.

[3] A medial moraine shown in the Boundary Survey photograph swings over to the west side; but we have no knowledge of its origin or relationships.

Plate LXV

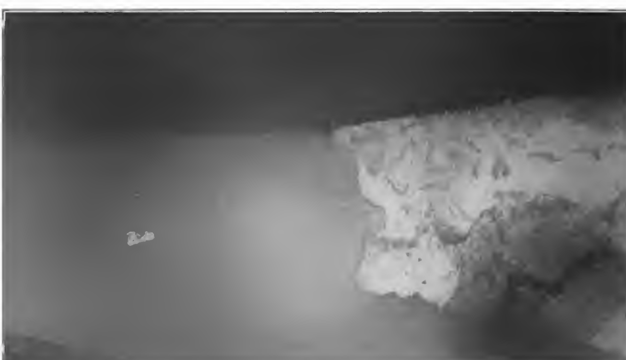
A.

B.

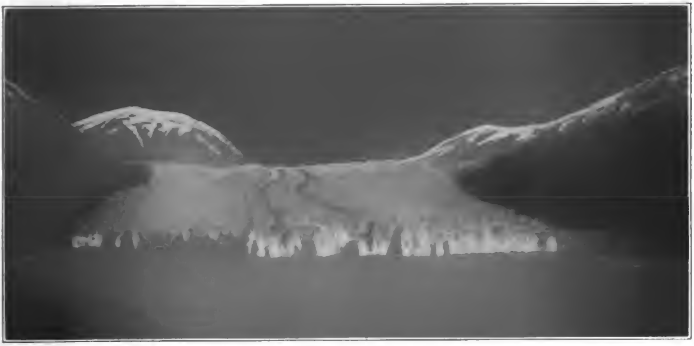
C.

Positions of North Edge of Tidal Arm of Nunatak Glacier in 1905, 1906 and 1909
From Station A (Map 4). See also Plate LXIV.

A. Front of Nunatak Glacier in 1905, from the Top of the Nunatak



B. Front of Nunatak Glacier in 1909, from the Top of the Nunatak



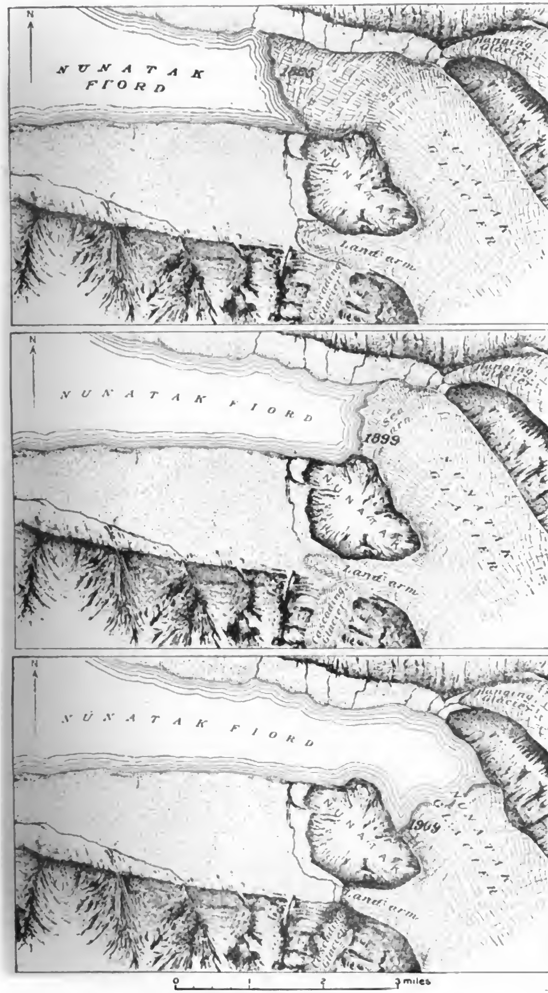
A. Nunatak Glacier in 1905
Photograph from Station G (Map 4).

B. Nunatak Glacier in 1909
From same point as upper view, showing recession in 4 years.

Plate LXVIII

Front of Nunatak Glacier in 1899
Photograph by D. G. Inverarity. Copyright, 1904, by E. H. Harriman.

PLATE LXIX



A.

B.

C.

NUNATAK GLACIER IN 1895, 1899, AND 1909 (TOPOGRAPHY AFTER GANNETT)

A. NUNATAK GLACIER IN 1909
Photograph from Station H (Map 4).

B. NUNATAK GLACIER IN 1910
Photograph from exactly the same point as upper view, showing advance in less than one year.

View Looking up a Broad, Glaciated, Hanging Valley

Glacier at its head descending to a moraine-covered end, which appears to be a low, dark hill in the center of view. The stream from the glacier flows in a shallow gorge on the right lower edge of view. Note the rock in the foreground and the broad U shape of the valley. Photograph taken July 5, 1906.

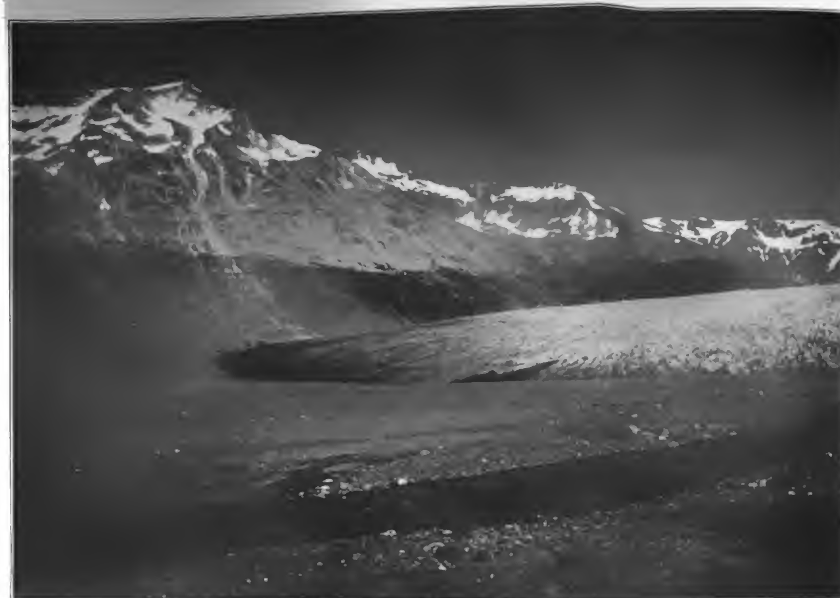
A. Cascading Glacier from the Nunatak

Showing hanging-valley conditions, the cascading of ice from ledge to ledge, and the ice sculpturing of the steepened valley slope, down which the water from the melting ice runs in numerous courses on the face of the rock (dark bands) without forming valleys. Photograph taken July 24, 1905.

B. Front of Hidden Glacier in 1899

Outwash plain in foreground. Photograph by G. K. Gilbert.

Plate LXXIII



Front of Hidden Glacier, North End in 1905

A. Ice Cave near Southern Edge of Hidden Glacier
From which main stream issued July 25, 1905.

B. Glacial Stream on South Side of Valley Train of Hidden Glacier in 1905

A. THE OUTWASH PLAIN OF HIDDEN GLACIER IN 1899
Photograph from Station C (Map 5), by G. K. Gilbert.

B. VIEW LOOKING NORTH ALONG THE FOSSE
Hidden Glacier on right; low gravel-veneered ice cliff on left, with several caves through which drainage escapes. The glacier is continuous across the fosse to the cliff. Overridden gravel terraces in background at mountain base. Photograph taken July 26, 1905.

Plate LXXVI

A. End of Hidden Glacier
From gravel terrace on north side of valley, June, 1899. Photograph by G. K. Gilbert.

B. End of Hidden Glacier in 1905
Note pronounced recession. Glacier terminus barely shows on extreme left. Northern stream in foreground, hanging valley on south side of valley, glacier end just to left of fosse, and gravel plain in front on site of glacier end in 1899.

A. North Edge of Hidden Glacier Resting on Older Ice-Eroded Gravels
The gash in the foreground is formed by a stream from the ice. Photograph taken July 25, 1905.

B. The Formerly "Hidden Glacier"
What we saw when we rounded the point in 1909.

A. The Surface and End of Hidden Glacier in 1909
Photographed from Station B (Map 5), after the two-mile advance.

B. Icebergs Massed on West Shore of Yakutat Bay in 1909

A. HIDDEN GLACIER IN 1906
Photographed from Station A (Map 5), before the two-mile advance.

B. SURFACE OF HIDDEN GLACIER IN 1905

From the point of junction of the tributaries down to the glacier terminus the Fourth Glacier has a remarkably smooth surface over which one could travel with ease in any direction. In smoothness it reminds us of the Hidden and Variegated Glaciers in 1905.

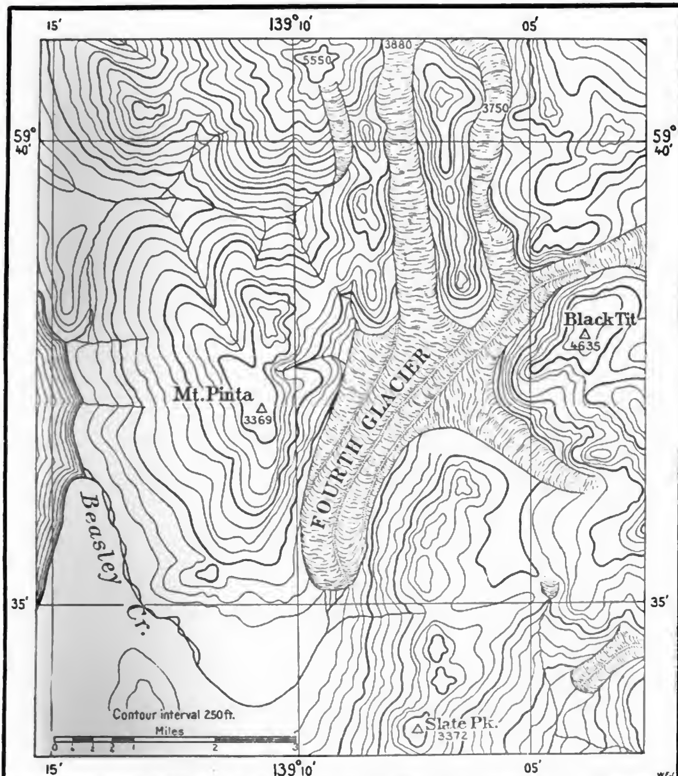


FIG. 15. FOURTH GLACIER IN 1906, BY CANADIAN BOUNDARY SURVEY.

It resembles Hidden Glacier also in the general absence of moraine cover, but in this respect it is widely different from the Variegated Glacier. Except for the lateral moraine on each margin of the glacier near the terminus, and the small medial moraine near the

east margin and a small area of terminal deposits described below, the ice surface is so free from débris that it appears clear in the photographs. Examined in detail one finds a small amount of moraine material scattered over the surface but not in quantity sufficient to form a continuous sheet.

The large proportion of clear ice in the Fourth Glacier, which resembles the condition of the Hidden, Nunatak, Orange, Hubbard, and Turner Glaciers, as contrasted to the débris-covered surfaces of such glaciers as the Butler, Haenke, Black, Galiano, Atrevida and Lucia, presents an interesting problem. Why should some of the glaciers have the condition illustrated in the Fourth Glacier and others the condition found in the Atrevida? In seeking to answer this question conclusively one would probably need to know more about the upper ice conditions in the larger glaciers, and whether the glaciers are actively moving, like the Hubbard, or slowly moving, like the uncrevassed Fourth Glacier. It is to be noted, however, that it is the larger glaciers that have the least amount of débris-cover. It is possible that the absence of débris is dependent upon the breadth of the valley, but under normal conditions the breadth of such a valley should diminish in the upper portions so that from the tributary sources, at least, such large quantities of débris would be supplied that it would form an ablation moraine over the wasting glacier end. The mere size of the valley in which the glacier ends does not, therefore, seem a sufficient explanation of the absence of ablation moraine in a region where so many glaciers have such extensive morainic cover. Only one hypothesis occurs to us which satisfactorily explains all the conditions, and this is that such glaciers as the Fourth, Hidden, and Orange, although large, receive their main snow and débris supply, not from the union of numerous small tributaries decending through narrow mountain valleys, but from snowfall upon broad ice divides and the avalanching of snow and rock on the mountain sides which enclose these through glaciers, with, of course, the addition of some small, short, relatively ineffective tributaries of the cascading type. By this explanation it would seem impossible for débris to extend in large quantities far enough out over the ice to furnish the material for a continuous cover of ablation moraine, while in such narrow valley glaciers as the Atrevida, débris would find its way out even to the center of the glacier. The absence of débris-cover on such active, clear-ice glaciers as the Hubbard, to which many narrow valley glacier tributaries doubtless contribute ice, may be due to the failure of ablation to lower the broken surface far enough to concentrate the débris by the time the iceberg-discharging sea cliff is reached.

Three facts indicate that the Fourth Glacier is not now in a very active stage. These are, first, its smoothness, second the fact that its front does not extend to the mouth of the mountain valley, and third, the evidence that it is now rapidly receding. Its front lies well within the mouth of the mountain valley, and there is, therefore, no opportunity for its expansion to form a piedmont ice bulb. In this respect Fourth Glacier differs widely from the Atrevida, Lucia, and Marvine and other tributaries of the Malaspina Glacier.

From the front of the Fourth Glacier two small streams emerge, one from either side. These flow along the front and join a large stream which issues near the center. This medial stream course is determined by a depression between the steeply-sloping alluvial fans which were constructed during a recent stage when the marginal drainage pursued a course down each side of the valley. In the depression Beasley Creek is building a small alluvial fan. It then flows along the east side of the mountain valley to the mouth

and turns westward, finally entering the head of Russell Fiord. A similar series of stream channels, now abandoned, show where an outlet formerly went down the west side of the valley, and between that course and Beasley Creek is a high wooded hill of dissected outwash gravels, rising to considerable height on the end nearest the glacier front (Pl. LXXXVIII).

The clear ice of the glacier projects farthest in the center of the valley, at the point where the stream passes down the depression between the bordering alluvial fans. On the very outermost portion of this point is some débris, staining the surface and forming débris cones and ridges. Beyond this, in front of the apparent end of the glacier, is a border of hummocky, débris-covered ice (Pl. LXXXVII) with many hollows in which water stands. Kettles, recent cracks, and pools of cold water show that the buried ice extends between seven and eight hundred feet in front of the centre of the apparent end of the glacier. The débris-covering of this outlying portion of the glacier is in large part gravel, laid down by deposit from the glacial streams. Débris-covered ice also projects beyond the front on both the margins of the glacier, where the lateral moraines give a covering which has checked recession. It is possible that ice also exists beneath the alluvial fans on the two margins of the valley, but no evidence of this was discovered.

The débris-covered marginal and frontal portions just described form a part of the evidence of recent notable recession. Absence of vegetation both on the mountain side above the glacier and in the flat in front of it, give further evidence of prolonged recession. The barren zone above the ice extends to a height of from 50 to 100 feet. Above this are found alder thickets, and on the west side at a still greater elevation, above 300 feet, there is spruce growth. This spruce is mature and it therefore appears that, although the glacier is now receding and has been doing so for some time, it has not expanded more than 300 feet above its present level for many years. At a distance of something more than a quarter of a mile from the glacier front, alder five or ten years old is growing, and cottonwood and spruce trees have advanced to within half a mile of the ice front.

These facts prove that the Fourth Glacier has been slowly receding for a considerable time, and that it has not been so greatly expanded as to reach beyond its mountain valley for many years, probably for more than a half-century. That it was, at some earlier stage, far more extensive than now, is made evident by the steepness of its valley walls and the presence of eroded spurs and numerous hanging valleys with lips from 500 to 1500 feet above the valley bottoms. The steepened slope and hanging valleys extend down beyond the end of the glacier even to the valley mouth. We may, therefore, safely infer that during an earlier stage, probably when Russell Fiord Glacier extended out on the foreland, and Yakutat Bay Glacier to the sea, the Fourth Glacier reached well out beyond its mountain valley and expanded in a piedmont ice bulb which perhaps coalesced with the Russell Fiord Glacier. One piece of evidence bearing upon this question of former extension of Fourth Glacier, is the fact that while the glacier is not now bringing crystalline rocks, but only sedimentaries from the Yakutat group, the gravels and bowlders in the stream bed beyond the mountain front include many crystalline rocks. The inference from this fact is that during the earlier expanded stage of the Fourth Glacier it received ice supply from a more distant source than now, in a region where crystalline rocks form the mountain walls. This source may have been at least as far back as the upper reaches of the Hidden Glacier. In all probability during this stage, the direction of ice flow in

this system of through valleys was different from the present, and it is possible that the breadth and flatness of the high divides of the present day are in large part the result of the flow of ice across the divides from which the glaciers now descend in two directions.

Although we had not visited the Fourth Glacier before 1909, we are convinced from its present condition that it has not yet been subjected to a forward thrust as a result of the influence of the 1899 earthquakes. Had it been, its surface would have necessarily retained some evidence of the former breaking, for it is inconceivable that ablation could have so reduced the irregularities as to have made such a smooth surface as the present. Four years of ablation on Variegated Glacier have made it possible to again travel over the surface, but has still left the surface so rough that one must use care in such traveling. Granting even six or seven years of ablation, which is the maximum which could possibly be inferred, a badly-broken ice surface could not be reduced to such smoothness as was observed on the Fourth Glacier. Further proof that this glacier has not been subjected to a spasmodic advance of considerable proportions is found in the fact that the glacier front and margins are fringed by extensive areas from which the glacier has receded. With recent advance under earthquake impulse evidence of the reverse condition should be present (Pl. LXXXVIII). We feel confident, therefore, that Fourth Glacier has not yet responded to the influence of the earthquake shocks. It was not visited in 1910 but its stream delta seemed unchanged at that time. Prospectors used it as a highway in 1898 and 1899 and in several years since, and we have heard no report of its being crevassed and impassable during the last ten years from these men, some of whose sledges and snowshoes we found on the glacier surface in 1909.

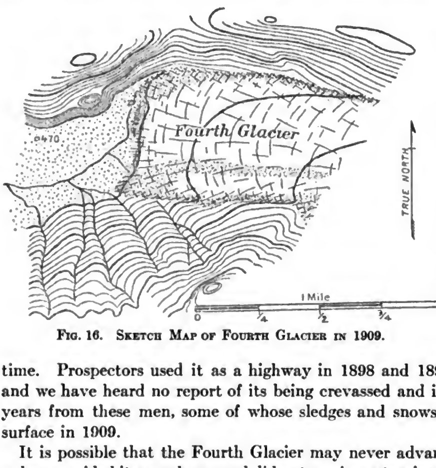


FIG. 16. SKETCH MAP OF FOURTH GLACIER IN 1909.

It is possible that the Fourth Glacier may never advance under the earthquake impulse, provided its supply ground did not receive extensive avalanches in 1899, but about this we cannot be certain. However, from such indications as we have concerning the conditions under which Fourth Glacier is supplied, we infer that it will ultimately respond to the impulse. It is, therefore, one of the glaciers which should be examined in future expeditions to the Yakutat Bay region. Two photographic stations were established in 1909, from which future observations may be made, and a sketch map of the ice front and its surroundings was made (Fig. 16). This, however, should be understood to be merely a record of general forms, the contours not being determined carefully, and all being located with respect to a single aneroid determination of altitude.

Smaller Glaciers

Besides the glaciers described in the preceding chapters there are a large number of minor glaciers in the Yakutat Bay region, many of which have as yet received no names. Some of the larger of these were named and briefly described in the reports on our 1905 and 1906 expeditions.[1] It does not seem important to repeat these descriptions here, though these glaciers present many interesting features, in detail, one of the most noteworthy being the broad extension of sheets of ablation moraine over their lower surfaces. Many of them resemble the Galiano and Atrevida Glaciers in this respect, though they lack expanded piedmont bulbs, since they no longer extend beyond the mountain front. Some, however, reach far down the mountain valleys, and one or two almost to the valley mouth, ending in low, undulating, moraine-covered surfaces, fringed by barren areas indicative of recent recession.

In 1909, 1910, and 1913 we made no additional observations upon these glaciers except to look at some of them from a distance. From this second view, and from a reconsideration of observations and photographs made in 1905, and in the light of the remarkable changes in some of the other glaciers observed in 1906 and 1909, we have concluded that there is reason for believing that some of these smaller glaciers had been subjected to advance and breaking between 1899 and 1905 as a result of earthquake shaking. Indications of this are best seen in the three glaciers on the west side of upper Russell Fiord, namely McCarty, Hendrickson and Rasmussen Glaciers. In 1905 the lower portions of the last two of these glaciers were completely covered, from side to side, with ablation moraine, having a very rough, hummocky surface. The McCarty Glacier was moderately crevassed but clean in its upper and middle portions (Pl. LXXXIX, B). Not then knowing the signs of recent advance, which we have since learned to interpret, we did not see, in the conditions of these glaciers, proof of recent advance and breaking. It is to be noted also that we would not have detected evidence of this even in the Galiano Glacier were it not for the fact that we had for comparison Russell's photographs and descriptions of the condition in 1891.

Re-examining the photographs of the small glaciers mentioned, we find the ablation moraine surface to be rough and angular, as if recently broken. It is noted also that there is complete absence of vegetation upon these moraine surfaces. This fact is perhaps the best evidence of recent advance, for the moraine is, in places, thick enough to support a growth of vegetation, and there is abundant alder growth on the hillsides above the glaciers. Its absence, therefore, does not seem possible of explanation either on the theory of failure of vegetation to advance or on the theory of prevention of growth through undermining and slumping. From distant views of these glaciers in 1909, we convinced ourselves that the ablation moraine was less rough than in 1905.

From this reconsideration of the evidence, in the light of subsequent discovery of the nature of phenomena to be expected as a result of a spasmodic advance of glaciers, we are inclined to believe that Galiano Glacier was not the only small glacier in this region to respond to the earthquake impulse prior to 1905. How many glaciers did advance in that interval will probably never be known, but it seems probable that many of the shorter glaciers were subjected to an advance and breaking within a few years after

[1] Tarr, R. S. and Martin, Lawrence, Glaciers and Glaciation of Yakutat Bay, Alaska, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 151, 152; Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 65–68.

the earthquakes shook down avalanches into their reservoirs. Among others we especially suspect that there was an advance of Blossom Island Glacier whose outer, moraine-covered surface was very rough in 1906 and whose front extended far beyond the mountain valley, and of Hendrickson Glacier which had moraine-covered ice resting upon overridden gravels in 1905, the moraine being entirely free from vegetation. McCarty and Rasmussen Glaciers may also have advanced before 1905.

Fallen Glacier

One of the most spectacular events in the recent glacial history of the Yakutat Bay region was the downfall of one of three small glaciers perched in steeply-sloping valleys on the west side of Disenchantment Bay. These three glaciers were so steeply perched in their hanging valleys as to attract our attention, and we were led to photograph them from the fiord on the 3rd of July, 1905, as had been done in 1890 by Russell, in 1895 by the boundary surveyors, and in 1899 by Gilbert. It happened that this day, July 3, 1905, was the last day in the existence of one of these glaciers, the southermost, which we now call Fallen Glacier. This glacier is estimated to have had a length of approximately a mile, and was supplied chiefly from a steep mountain crest down which the snow slid into a spoon-shaped cirque about halfway down the slope. From this amphitheatre the ice protruded as a crevassed, cascading tongue, the lower end lying at an elevation of about 1000 feet above the fiord, and separated from it by an ice-steepened rock slope. On the 4th of July, the day after we photographed the glacier, the entire glacier mass slid out of its valley and a large part of it tumbled into the fiord. The fall of so much ice into the water started a water wave which rose to a height of 110 feet on the shore just south of the glacier, breaking off alder bushes at that elevation. Three miles to the north, near Turner Glacier, the water wave killed vegetation at an elevation of 55 feet, and a wave swept across the north end of Haenke Island at an elevation of 50 or 60 feet. At that point in one place where it was locally concentrated, the wave washed out good-sized alders at an elevation of 115 feet.

In addition to the breaking off and uprooting of alder bushes, the wave caused much erosion wherever it came into contact with unconsolidated deposits, cutting small cliffs and gullies in the till and alluvial fans. The annual plants were killed up to the elevation which the water wave reached, and in succeeding days it was possible to trace the level to which the wave rose by the zone of parched brown grass and other vegetation killed by the bath in salt water. Small icebergs were left stranded far above high tide in several places, a number being found on the site where our camp had stood on the day before the fall of the glacier. We had fortunately moved our camp a distance of about 15 miles, to a new site in Russell Fiord, where we were separated from the glacier fall by Osier Island, and by the high mountainous point above it.

At the time of the glacier fall we were working along the coast east of our camp and were surprised by the appearance of a series of waves, which at first were ascribed to some unusually large fall of icebergs from Hubbard Glacier; but since the waves increased in height until they finally rose 15 or 20 feet above the water level, and since they were much higher than any previously observed iceberg waves, and lasted much longer, we inferred that they must be the result of an earthquake, until the next day, when one of our native packers, returning from Yakutat, gave us the true explanation of the phenomenon. The native further stated that this was the third time that this glacier had

fallen from its valley, though on questioning him it was evident that the tradition referred merely to the falling of a glacier from its valley on the west side of Disenchantment Bay, not surely this one each time. The last fall, which he said occurred about sixty years before, he reported to have destroyed 100 natives who at the time were encamped at their summer sealing camp a few miles south of Haenke Island. Fortunately at the time of the 1905 fall the natives had all left the bay.

By the avalanching of this glacier, its valley was completely emptied of ice, with the exception of a small remnant of steeply-perched névé and a few minor ice fragments near the edge of the cirque. The walls and bottom of the cirque were of bare rock, and the avalanche, which spread out fan-shaped at the mountain base, had swept away the soil, and for a width of half a mile had killed the alder growth which previously grew at the cliff base beneath the glacier. The larger part of the glacier fell into the fiord, probably most of the ice floated away, while the loose débris sunk to the bottom. Some of the glacier ice remained at the base of the cliff, pushing the coastline outward slightly and forming a new shore line of angular rock-débris. That ice existed beneath this was evident from the fresh faulting and slumping of the surface in 1905.

In 1909 we again crossed this area and found that ice still existed beneath the débris that was swept out of the mountain valley in 1905. The cirque valley of Fallen Glacier is beginning to be again filled with snow, and the place of the former glacier is now covered by one of the largest snowfields on this mountain face (Pl. LXIX). There has not yet been sufficient accummulation of snow to cause the redevelopment of the glacier, but we have here the early stages of formation of a new, small, perched glacier which may perhaps be reaccummulating to once more avalanche out of its valley, when a sufficient amount has gathered here to again render its position unstable.

CHAPTER X

THE EARTHQUAKE ADVANCE THEORY

The Problem. The facts stated in the previous chapters make it clear that the glaciers of Yakutat Bay have in recent years been undergoing a series of changes of unusual character. From a state of stagnation a number of them have sprung into sudden activity and then, with almost equal abruptness, have relapsed again into a state of stagnation. The transformation has affected different glaciers at different times and in different degree, and some of the glaciers have not been affected at all. So far as we know, such spectacular changes have not hitherto been recorded. The advance of Vernagt-Ferner in the Tyrolese Alps is a case analogous to these, but is only a single instance in a glaciated region, and it differs in notable respects from the advancing Yakutat Bay glaciers. The advance of two glaciers in Icy Fiord, Spitzbergen, and the advance of glaciers reported from the Himalayas are too little known to offer basis for comparison. The solution of the problem of the cause for these changes must, therefore, be based mainly upon the facts which the Yakutat Bay region presents.

Upon the basis of the comparative observations of 1905 and 1906 the senior author has proposed the theory of avalanching during earthquake shaking, rather than increase of snowfall with climatic oscillations, as the explanation of the phenomena of advancing glaciers in this region.[1] The observations of 1909, 1910, and 1913 add important and significant facts to those previously observed, and among these facts are several tending toward the verification of the theory, and none opposed to it. With these new facts we are in a position to more fully discuss this theory than was hitherto possible. This discussion will be prefaced by a summary of the significant facts of observation which have already been presented in some detail under the description of the individual glaciers.

Summary of the Phenomena. In recent years, up to Gilbert's visit in 1899, the general history of the Yakutat Bay glaciers had been one of recession; and in most of them the recession continued to 1905, and in some even to 1913. This recession was apparently a stage in the withdrawal of the glaciers from positions to which they had readvanced after a period of recession even greater than the present. During this earlier advance Nunatak Glacier had pushed northwestward into the lower portion of Russell Fiord, and also southward up the fiord until it joined the Hidden Glacier and the united ice stream advanced far up toward the head of the fiord. From this advanced position the recession of the glaciers has been both rapid and recent, and in the case of the Hidden Glacier has continued until 1906 or 1907, and in the Nunatak Glacier until 1909. It is not certain that this recession was not interrupted by halts or minor advances; but there is no evidence of this, and certainly, during the period of observation, that is

[1] Tarr, R. S., Recent Advance of Glaciers in the Yakutat Bay Region, Alaska, Bull. Geol. Soc. Amer., Vol. 18, 1907, pp. 277–286; The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 90–95; The Theory of Advance of Glaciers in Response to Earthquake Shaking, Zeitschrift für Gletscherkunde, Band V, 1910, pp. 1-35; Tarr, R. S. and Martin, Lawrence, Professional Paper 69, U. S. Geol. Survey, 1912, pp. 51-61.

since 1899, there has been no interruption as great as the advance of the Hidden Glacier in 1906 or 1907.

In 1905 evidence was found to prove that Galiano Glacier had undergone great changes since last observed and photographed in 1891. Its moraine-covered surface had been so broken as to completely destroy an alder and cottonwood thicket that grew on it in 1890 and 1891, but ablation had healed the broken surface so that one could easily travel over it in 1905. There is some evidence that the front of the ice bulb advanced, and sound basis for the inference that there was lateral spreading toward the valley walls, destroying marginal drainage. The changes in the glacier condition extended to a hitherto undiscovered, stagnant, off-lying ice mass on which an alluvial fan had been deposited, and by the advance this fan was destroyed and its place in large part taken by morainic mounds thrust up through it. We have no means of positively determining in which of the fourteen years between 1891 and 1905 these changes occurred but photographs make it certain that it was after 1895, and we infer that it was in 1900 or 1901 because of the age of the alder bushes that have again begun to grow on the moraine. Other facts indicate recency of the change, such for example as the imperfect redevelopment of marginal drainage, and of the alluvial fan between the morainic knolls. Furthermore, we are at a loss to explain the transformation of this glacier on any other hypothesis than the effect of earthquake shaking, and are forced to the conclusion that the advance is the result of a response to the shaking during the earthquakes in September, 1899. The Galiano Glacier, whose response was so rapid, is a very short glacier, with steep valley sides and head, on which in September deep snows had already been added to the extensive snowfields and steeply-descending glaciers. Evidence that there has been extensive avalanching from these steep slopes was readily discovered by a comparison of the 1891 photographs with the condition in 1905; and even in 1909 the spaces from which large masses of ice had fallen after 1891 were not yet reoccupied by ice.

It seems probable, though the facts do not conclusively prove it, that other small glaciers, similar to Galiano Glacier, had undergone a like change a short time before the expedition of 1905. Had we then had suspicion of the possibility of such glacier transformation we might have discovered evidence which is now obscured by ablation. The glaciers which gave the best indication of having undergone changes similar to those of the Galiano are McCarty, Hendrickson and Rasmussen Glaciers on the west side of Russell Fiord, and the Blossom Island Glacier just north of Blossom Island east of Marvine Glacier. It is probable that some of the small glaciers have failed to show evidence of notable response to earthquake shaking because of the small supply of snow and ice thrown down upon them. Of one such glacier, the Black Glacier, the one nearest the Galiano, the evidence is convincing that there has been no distinct advance of the end, no pronounced lateral spreading, nor sufficient breaking of the surface to destroy the alder growth upon it. Not having yet advanced as a result of the 1899 earthquakes, we assume that this and other similar glaciers will probably not do so in the future.

In 1906 four glaciers, the Haenke, Atrevida, Variegated, and Marvine, had become absolutely transformed from their condition in 1905. Three of these glaciers are of about the same order, and it seems natural that they should advance essentially together. Moreover, their size is such in comparison with the Galiano that it is not difficult to understand why they should require so much longer to respond to the earthquake

shaking. The Marvine Glacier, on the other hand, seems to be much larger than either of these, and certainly is if the Malaspina extension of Marvine Glacier is included. We do not know what the conditions are in its upper portion, and it is possible that the advance is the result of a thrust from some of the shorter tributaries; or it may be true that this glacier is a steep one, though broad and flat where it emerges from its mountain valley, expanding far beyond the mountains because of large supply grounds at no great distance back in the mountains. The best map of the region that we have indicates that the latter is true, and that the Marvine Glacier is little if any longer than the Variegated Glacier and much shorter than the Lucia Glacier. If this is not true, it is certainly noteworthy that this glacier should have advanced as early as the Atrevida, and three years before the Lucia did. If we could depend absolutely on the map the question would be easily settled.

By the advance of these four glaciers the ice surface was in each case profoundly broken, being transformed from a surface over which one could easily travel in 1905 to one which was utterly impassable in 1906. In at least two cases, the Atrevida and Variegated Glaciers, there was a lateral spreading of the ice within the mountain valley, closing up marginal valleys which had previously existed. In these two cases, and probably in the others also, there was a pronounced thickening of the glacier at and beyond the mountain valley mouth. Three of the glaciers—the Marvine, Atrevida and Variegated—terminate in expanded piedmont bulbs, and in each of them the breaking of the ice surface extended into these ice bulbs, in the Marvine reaching to the very outer edge, but in the two others failing to affect the outermost parts of the stagnant ice. From this we infer that the thrust in the Marvine Glacier was the most pronounced of all and this inference is supported by the further fact that the extent of the area of broken ice was far greater in the Marvine than in the other glaciers. The area of broken ice visible in the Marvine Glacier, and its continuation in the eastern lobe of Malaspina Glacier, was fully 15 miles in length and from 8 to 10 miles wide in the broadest part; in Atrevida and Variegated Glaciers the broken ice occupied an area whose length was between 5 and 7 miles, with a breadth of 2 or 3 miles in the broadest part. All four of the glaciers advanced along their fronts and margins, the greatest advance being in the Haenke Glacier, the only one whose terminus was confined between mountain walls. Here the front of the glacier was pushed forward nearly a mile and it was transformed from non-tidal to tidal condition by the thrust. The forward motion in the other three glaciers was partly dissipated in breaking the piedmont area and partly in thickening the ice in it; but some of the thrust was effective in pushing the margins of the piedmont areas outward. This was clearly shown in Variegated Glacier by the overriding of a granite gorge; in the others by the advance of the margin into the fringing forest.

Thus, any theory which will account for the transformation of these glaciers, must explain not only change from stagnant condition to activity, and from smooth to broken surface, but also actual forward movement and pronounced thickening of the glaciers. The theory must also account for the suddenness of the change. At least three of the glaciers, Atrevida, Haenke, and Variegated, and probably also a fourth (the Marvine) passed through the complete cycle in approximately one year. In the last of August, 1905, they showed almost no evidence of change, though in the Atrevida and Marvine Glaciers there were observed phenomena in 1905 that were later interpreted as the beginning of the advance; in June, 1906, all four glaciers were utterly transformed. The

change was as spasmodic as it was absolute, and we know of no better phrase with which to characterize it than to call it a *glacier flood*. Not only was the advance abrupt, but its termination was equally so, for it ceased with the season of 1906, and since then ablation has been at work healing the broken ice which has again relapsed into a stage of stagnation. The theory must, therefore, account for an abrupt and absolute cessation of the advance as well as for the spasmodic advance itself.

Between 1899 and 1901, and also between 1906 and 1909, there have been slight fluctuations in the fronts of Turner and Hubbard Glaciers, which may be due to thrusts from tributaries; but there has been no notable advance of these great tidal glaciers, although in 1909 Hubbard Glacier showed signs of the beginning of an advance which in 1909 we thought might prove to be of importance. Since, however, it was not followed in 1910 by a distinct forward movement we infer that it is not an advance comparable with those of the other advancing glaciers. Between 1906 and 1909 only one glacier, the Hidden, suffered advance, probably in the season of 1907. When last seen in early July, 1906, its terminus was in the same smooth, stagnant condition as in 1899 and 1905; but in early July, 1909, its front was about two miles farther down its valley. There had been accompanying spreading on the margins and a thickening of at least eleven hundred feet where the glacier terminus stood in 1905. With the advance came profound breaking of the surface, and, succeeding it, ablation in sufficient degree to render the surface passable. Thus the Hidden Glacier has repeated the phenomena of the other advancing glaciers, but the advance of its front has been greater than that of any other. Its change from stagnant to active condition and back again to stagnation was abrupt and all compassed in a short time, probably not over a year.

In 1905 and 1906 Lucia Glacier showed no sign of past change, but in 1909 its moraine-covered lower portion, and also the valley glacier above, was greatly crevassed, there was a lateral spreading in progress, and a distributary on the western side had advanced notably, while the ice had surrounded and piled upon the northern end of the nunatak near by. There was evidence that the advance and breaking had only recently begun and was still in progress. The surface was not so broken as the other glaciers, and the effect of the advance had not extended so far out in the stagnant piedmont bulb as in the case of the neighboring and contiguous Atrevida Glacier in 1906. Hubbard Glacier in 1909 showed an even earlier stage in advance than Lucia Glacier. Its eastern margin was newly broken, but what had occurred farther out in the glacier could not be told because the surface of this active glacier has been severely crevassed, except at the very margin, during the entire period of observation. However, the front of the glacier had moved forward somewhat since 1906 and there was evidence that the north arm had become more powerful. There was little additional advance in this arm between 1909 and 1910, though the west part of the glacier had continued the slower advance.

In 1910 the Nunatak Glacier began to advance and up to June its tidal front had come out 700 to 1000 feet with some thickening, but, because confined in a narrow fiord, with practically no lateral spreading. This seems to have been the whole of the advance and is probably to be ascribed to a response to a thrust from a single small tributary.

TABULAR SUMMARY OF ADVANCES SINCE 1899 EARTHQUAKES

Year of Advance	*Name of Glacier*	*Length of Glacier*
After 1895 and before 1905	Galiano	2 or 3 miles
Between 1899 and 1901	Miller	3 or 4 miles
	[1]	
	[2]	
1905–1906	Haenke	6 or 7 miles
	Atrevida	8 miles
	Variegated	10 miles
	Marvine	10 miles (excluding expanded lobe in Malaspina)
1906–1907	Hidden	16–17 miles
	[3]	
	[4]	
1909	Lucia	17–18 miles
	[5]	
1910	Nunatak	20 miles

[1] Perhaps Hendrickson, Rasmussen, McCarty, and Blossom Island Glaciers, each 1 to 2½ miles long.
[2] Perhaps Turner between 1899 and 1901, and Hubbard between 1899 and 1901, and Hubbard between 1901 and 1905.
[3] Perhaps Turner between 1906 and 1909.
[4] Northwestern arm of Hubbard.
[5] South arm of Hubbard.

Generalizing the facts observed in the various glaciers, during our five seasons of observation, it may be said that certain glaciers, nine certainly and perhaps six others, present evidence of an abrupt and absolute transformation within a short time, and then, in every case, a sudden return to their former condition. Among the glaciers that have advanced there is a general agreement between size of the glacier and time of the advance, as shown in the tabular statement above. The Galiano, which advanced before 1905, is one of the smallest visible glaciers of the region and Miller Glacier, which advanced between 1899 and 1901, is little if any larger; the four that advanced in 1905–6 are larger than the Galiano and Miller, though not among the largest Yakutat Bay glaciers; the Hidden Glacier, which advanced later, probably in 1907, is still larger; the Lucia, which advanced in 1909, is larger than the Hidden; and the Nunatak, which advanced in 1910, is larger than the Lucia. The one apparent exception to this order of appearance of the advance is the Marvine-Malaspina Glacier, whose valley portion falls in its proper place in order of size, if we may accept the map as correct. Two of the largest and most active glaciers (Turner and Hubbard) have had no notable advance, though there have been some fluctuations of their fronts, perhaps representing response to thrusts from shorter tributaries. For some of the smaller glaciers, we have no proof of an advance; but, in the case of some of the smaller of these, this may be due to our failure to recognize evidence of an advance that was completed before 1905; in other small gla-

ciers, and certainly in one, Black Glacier, there has been no advance, and probably will be none because of unfavorable conditions in the reservoirs. The fact that some of the larger glaciers, like the Seward and Hubbard, have not yet undergone transformation can hardly be due to this cause; and it is reasonably safe to predict a change for them also, when time enough has elapsed for the impulse to pass from their distant reservoirs, through the long valley ice streams to their fronts. When one of the large tidal glaciers responds the series will be complete for this region.

Elsewhere in Alaska there has been no analogous transformation of glaciers, so far as can be learned. Such inquiry as we have been able to make reveals no evidence of recent notable change in the glaciers of southeastern Alaska with the exception of Glacier Bay; and our studies to the northwest of it, in the Copper River and Prince William Sound regions, and those of Professor Grant in the latter region and the Kenai Peninsula, fail to discover evidence of changes of like character in the glaciers of these sections. As is stated in the chapters describing our observations in Prince William Sound, there have been some recent changes in several of the glaciers, but nothing that compares with the remarkable changes in the Yakutat Bay region. We feel warranted, therefore, in concluding that the spasmodic advance and absolute transformation of glaciers during the past 15 years is a phenomenon of Yakutat Bay and vicinity. That past changes of similar character may have occurred in other parts of Alaska seems probable; and it is believed possible that certain advances noticed in the Glacier Bay region, and elsewhere in Alaska, are similar in origin to the advance of the Yakutat Bay glaciers but with less intense cause, and consequently with less spectacular results.

An examination of the records of changes in glaciers in other regions has failed to discover either transformation or spasmodic advance by a group of glaciers of like nature to the phenomena of Yakutat Bay. There are many instances of forward movements succeeding periods of recession, and some of them have been spasmodic in their nature; but so far we have been unable to find anything comparable to the changes described in the preceding chapters. In this respect the glaciers of Yakutat Bay are unique among the glaciers of the world, so far as the literature which we have seen goes to show.

It is noteworthy, also, that in September, 1899, the Yakutat Bay region was the center of a series of earthquakes of great intensity; two or three of them ranking with the greatest of recent earthquakes.[1] For three weeks the region was repeatedly shaken, and there were hundreds of shocks in all. In our reports upon these earthquakes abundant evidence has been presented to prove that during the earthquakes there were frequent and enormous avalanches of snow, ice, and rock. These avalanches have not only been testified to by prospectors encamped within Disenchantment Bay during the earthquakes, who saw and heard them fall, by the natives who report that certain mountain faces were "completely changed" by the earthquakes, but by our own observation of exceedingly abundant, fresh avalanche scars upon the mountains, far more abundant than in other mountain regions of Alaska that were not severely shaken during these earthquakes. Comparison of photographs taken before the 1899 earthquakes with those taken after it show conclusively what large areas of avalanches there were in those places.

[1] Tarr, R. S. and Martin, Lawrence, Recent Changes of Level in the Yakutat Bay Region, Alaska, Bull. Geol. Soc. Amer., Vol. 17, 1906, pp. 29–64; Geog. Journ., Vol. XXVIII, 1906, pp. 30–42; The Earthquakes at Yakutat Bay, Alaska, in September, 1899, Professional Paper 69, U. S. Geol. Survey, 1912; Martin, Lawrence, The Alaskan Earthquakes of 1899, Bull. Geol. Soc. Amer., Vol. XXI, 1910, pp. 339–406.

In other parts of the St. Elias Range, geologists, army officers, prospectors, and others saw and heard many great avalanches during the earthquakes, although some of these observers were several hundred miles from the earthquake centre; but in these remote regions the avalanches were not in such abundance as near Yakutat Bay.

One who has read Russell's vivid descriptions of the great avalanches in the valley of upper Newton glacier[1] in 1890 and 1891, some of which advanced a half mile across the glacier, or who has seen Sella's wonderful photographs[2] of the avalanche-scarred slopes near the heads of the glaciers of Mt. St. Elias in 1897, in both cases representing avalanching under *normal* conditions, not under the exceptional conditions of three weeks of earthquake shocks, and who has noted the enormous area of snow slopes in unstable equilibrium near Yakutat Bay, could have no serious doubt of the tremendous accessions of snow and ice that might have been received by many of the Yakutat Bay glaciers during September, 1899. Upon these evidences, we are convinced that there can be no question but that there was, in that year unusual addition to the reservoirs of the glaciers of Yakutat Bay and vicinity. The mountains here are lofty, with steep slopes, and the snowfall is very heavy. The same statement applies to other parts of the mountains of the Alaskan coast region, it is true; but, although these other sections have experienced earthquakes, none have suffered such severe and continued shaking since Alaska began to be settled by white men as the Yakutat Bay region was subjected to in September, 1899. That period of shaking was notable for its intensity and duration, even in a region of frequent and severe earthquakes. If the avalanching of snow and ice to glacier reservoirs is competent to cause an advance in the glacier termini, the advance resulting from that particular period of shaking should be notable for its extent and intensity. Lesser shaking might well cause slower and less spectacular advances, of which the slighter forward movements of the glaciers of Alaska and other regions may be instances; but vigorous shaking, continued through three weeks, may well produce additions to the glacier supply that cannot be disposed of in so simple a manner.

Until the studies of the Yakutat Bay glaciers in 1906 the theory of advance of glaciers in response to earthquake shaking had not seemed called for in explanation of the phenomena of glacier fluctuations. From the studies of that year it became evident that some unusual cause was necessary to explain the remarkable changes observed in the glaciers, and the earthquake theory was therefore proposed. Our later studies have discovered new facts in support of this theory and none opposed to it. In view of the newness of this theory, and of its possible application to glacier advances in other regions, we deem it worth while to state it fully, to give a complete statement of the reasons why it seems to be necessary, and to state the facts which support it.

Failure of Other Theories. One of the main reasons for proposing the new theory of advance due to an impulse derived from earthquake shaking is the failure of all other explanations suggested. It was evident that we had here such a strong expression of some cause for glacier advance that we were warranted in considering not only current theories, but unusual and even apparently improbable theories, and in inventing for consideration any that might by any possibility explain the phenomena. Little or no help could be obtained from observations in other regions, because the phenomena

[1] Russell, I. C., Nat. Geog. Mag., Vol. III, 1891, pp. 155–156; 13th Ann. Rept., U. S. Geol. Survey, 1892, p. 50.
[2] The Ascent of Mt. St. Elias, London, 1900, photograph facing p. 130, and many others.

observed in Yakutat Bay were quite different in degree and in character from those of other regions.

Hypothesis of Climatic Variation. The consideration of the problem naturally began with the commonly-accepted theory for fluctuations in glacier fronts—response to variations in snowfall in the glacier reservoirs. No one questions the rationality of this theory when applied to such slow advances and recessions as occur, for instance, in the Alps, where for a series of years ice fronts advance or recede slightly; but it is a question worthy of careful consideration whether this cause is also competent to explain the spectacular changes observed in the Yakutat Bay region.

In considering this question it is to be borne in mind that we have to account for the sudden advance of nine or ten glaciers in a limited region, and that these glaciers vary in size and in the conditions of their reservoirs.

Under the theory of climatic variations it would be necessary to assume that after a period of a half century, or thereabouts, of snowfall not quite sufficient to hold glacier fronts in their position, thus giving rise to stagnation in their outer portions, and accompanying recession, the snowfall abruptly increased and enough fell to produce these startling phenomena. Then came a cessation of the excessive snowfall as abrupt as the beginning. In other words, in a limited area there must have been a great increase in snowfall for a few years and then a return to the normal. It must be insisted, too, that the phenomena of advance show clearly that the transformation could not be accounted for by moderate increase in snowfall. Only a very notable addition to the reservoirs could bring about so spasmodic and so great a forward movement.

In the hope of getting some light on the question of variation in snowfall in this region we have examined the precipitation data for stations on the Alaskan coast within a distance of 250 miles to the southeast and an equal distance to the northwest.[1] These records, though fragmentary, prove, what is otherwise well known, that there are great variations in amount of precipitation from place to place along this coast, and since the nearest of these stations is 150 miles distant from Yakutat Bay, it is evident that no safe deduction can be drawn from the recorded precipitation in an interpretation of variation in snowfall on the slopes of the St. Elias Range. We do not know whether the precipitation on the mountain slopes is greater or less than at Sitka or Nuchek, nor do we know how it varies with elevation or exposure, nor whether an increase in precipitation at Sitka was associated with an increase on the slopes of the St. Elias mountains. After a careful consideration of the question we have become convinced that no definite or satisfactory discussion of the subject is possible on the basis of existing records of precipitation. On one point, however, there is noteworthy testimony from the records, namely that, for some reason, there is, at least locally, a very decided variation in precipitation. If the records are accurate, there has been at Sitka, 250 miles southeast of Yakutat, a variation in rainfall of from 59 inches in 1872 to 140 inches in 1886. From 1848 to 1876 the average precipitation was 79 inches, and in no year did it rise above 94 inches, while in only two years did it fall below 65 inches. Up to this time the precipitation was fairly regular, but a precipitation of 77 inches in 1876 was followed by one

[1] Based upon records made by the Russians and published in House Ex. Doc. 177, 40th Congress, 2nd Session, Washington, 1868, pp. 334–5, 337; upon U. S. Weather Bureau records published by Cleveland Abbe, Jr., n The Geography and Geology of Alaska, Professional Paper 45, U. S. Geol. Survey, 1906, pp. 188–200; and upon later manuscript records furnished us by the U. S. Weather Bureau.

of 102 inches in 1882 and 1883, 110 inches in 1884, 102 inches in 1885, and 140 inches in 1886. From 1890 to 1899 inclusive there are no records; but from 1900 to 1909 the precipitation has been fairly regular, the highest being 88 inches, the lowest 73 inches.

At Killisnoo, 40 miles northeast of Sitka, records from 1890 to 1905 give a range between 43 inches and 72 inches. At Fort Liscum, near Valdez, 250 miles northwest of Yakutat Bay, records for seven years between 1901 and 1909 show a variation between 56 inches in 1904 and 91 inches in 1907. Other records are even more fragmentary. Indeed, some of the variations are so remarkable that one is almost tempted to question their accuracy, as when the Sitka precipitation the year before the temporary cessation of records (1886) leaped up to 140 inches, and the records at Chilkat, which though kept for only two years give 39 inches for 1884 and only 11 inches in 1885.[1]

The precipitation records, though showing remarkable variations at a given station, fail to show similar variations for the same years at other stations. Thus we must assume that whatever variations there are, in those years where comparisons are possible, are local, not general. Unfortunately this comparison cannot be carried very far, though it may be made in some cases. For example, the year (1885) when Chilkat had a precipitation of only 11 inches, Sitka had 102 inches! The highest precipitation at Sitka since 1900 is 88 inches (in 1901 and 1902), but in these same years the nearby station of Killisnoo had only 55 and 45 inches respectively, both below the average precipitation, which is 56 inches. The highest recorded precipitation at Fort Liscum is 91 inches, in 1907, but in that year Sitka had only 79 inches, or much less than the average.

It is a question whether one is warranted in drawing any definite conclusions from such fragmentary and variable statistics, but in the light of these records we cannot fail to admit at least the possibility of two conditions bearing upon the problem with which we are dealing: (1) that there may be very extensive variations in the amount of snowfall from place to place in the Alaskan coast mountains; (2) that there may be great local variations in the amount of precipitation there from year to year.

This complicates the solution of the problem of the cause of the advancing glaciers and introduces difficulty in the definite elimination of the climatic hypothesis. Yet, in spite of these possible great variations in snowfall, we believe the climatic hypothesis to be inadequate to account for the spasmodic advance of the Yakutat Bay glaciers. We may perhaps best state our reasons for this belief by first considering a single one of the Yakutat Bay glaciers in comparison with the best notable known instance of glacier advance correlated with variation in snowfall, the Vernagt Glacier of the Tyrolese Alps.[2] On several occasions this glacier has undergone spasmodic expansion with accompanying transformation from generally uncrevassed to crevassed condition, the last periods being in 1845 and 1898–1900.[3] Although a small glacier, the Vernagt is peculiar in having a very large reservoir with a narrow ice tongue extending from it. In this way the effect of increase in snowfall is magnified and a wave of advance passing down this

[1] On this point, however, Dr. Cleveland Abbe writes in a letter to us under date of May 30, 1910:—"I would express my belief that the reports of rainfall at Sitka for 1882 to 1886 must be largely influenced by some temporary peculiarity in methods or gauges, or measurements. I cannot consider the excessive figures, 102 to 104 inches, as homogenous with the rest of the column for Sitka; they are not mere natural irregularities of precipitation."

[2] Finsterwalder, S., Der Vernagt-Ferner, Wissenschaftliche Ergänzungshefte zur Zeitschrift des D. u. Ö Alpenvereins, I Band, 1 Heft, Graz, 1897.

[3] For reference, see footnote, p. 182.

glacier causes great extension of the glacier end. If we may assume this glacier to show a normal type of notable advance due to increase in snowfall, which might be compared with the Yakutat Bay glaciers, there is one very noteworthy difference. While the Yakutat Bay advance began and ended in a period of ten months, the Vernagt, in its last advance, moved forward much more slowly and through a much longer period of time. It was, in other words, less spasmodic. This, however, is not a vital point, for it is conceivable that with still greater increase in snowfall the advance might have been more spasmodic, as it apparently was in earlier periods when its terminus was pushed much farther down the valley than in 1899–1900.

Turning to the Yakutat Bay region, and selecting one of the glaciers there for comparison, we might almost equally well take either the Atrevida or the Galiano Glacier, but the latter will serve our purpose best because it is the smaller and therefore more nearly comparable with the Vernagt. Its advance was great, spasmodic, and effective. It differs from the Vernagt in having no great snowfield reservoir with which to intensify the effect of increased snowfall. To account for the advance, therefore, one would need to assume a far greater increase in snowfall than was required in the Vernagt, and exactly the same statements apply to the Atrevida Glacier. How much greater the snowfall would need to be we do not know, but surely to account for such changes as were observed would require a vast addition to the upper glacier, perhaps even more than the greatest recorded variation in precipitation in any of the Alaskan records, continued through several years.

Although such increase does not seem to us at all probable, we have to admit the possibility of it in view of the records of Alaskan precipitation. However, assuming such increase in precipitation, and further assuming that it was localized in the Yakutat Bay region, and that it was competent to account for the advance of Atrevida and Galiano Glaciers, we are at once confronted by the very serious difficulty that Black Glacier, heading in the same mountains as the Galiano, and only five or six miles away, has not responded. Its failure to advance as a result of snow supply added by earthquake shaking is explicable, but that it should not have responded to such a great increase of snowfall as would be required to account for the advance of Galiano and Atrevida Glaciers is inexplicable. The Black Glacier resembles these two glaciers in being of small size, in having no enlarged reservoir, in receiving its supply from snowfields and steeply perched tributaries, and in passing down a short, deep mountain valley enclosed between precipitous walls. It should show at least some response to greatly increased snowfall; it might fail to show response to earthquake shaking.

With these facts in mind we cannot retain the climatic hypothesis, even though granting the maximum variation in precipitation that anyone could possibly claim. But aside from this, the hypothesis as applied to this region is weak in other respects. It must assume a localized increase in snowfall of great amount, without any explanation either of its localization or of its cause, and without any proof of its occurrence. In this respect the climatic hypothesis compares unfavorably with the earthquake hypothesis, for by that we know why there was localization of effect, we know that vigorous and long continued earthquake shaking occurred, and we know that vast quantities of snow and ice were added to the glacier reservoirs. The climatic hypothesis is difficult to accept also because of the unusual, and in fact unparalleled, phenomena which it would be called upon to explain. While it is true that the Vernagt Glacier has advanced in a

similar way, it is nevertheless a special case and its advance under snowfall variation in due to peculiar individual characteristics. So far as known the Yakutat Bay region is the only one in the world where a number of glaciers of different forms, lengths and reservoir conditions have undergone a spasmodic and well graded advance and complete transformation. To account for this by climatic variation calls for a far greater increase in snowfall than has hitherto been recorded in glacier regions. Being unprecedented does not make it impossible, to be sure, but it adds one more difficulty in the way of acceptance of the hypothesis. We find it difficult also to harmonize the climatic hypothesis with the fact that the phenomena of spasmodic advance are, so far as we know, localized in and near the Yakutat Bay region. So great advance as is observed here would demand notable increase in snowfall for a period of certainly a few years; and we are unable to understand why there should be this great increase in one area, and nothing corresponding to it in the adjacent regions, where mountains rise equally high, where snowfields are equally extensive, and where glaciers are equally large. One difficulty in understanding this leads to skepticism when we remember that the region in question is the one which was most severely shaken by the earthquakes of September, 1899. In fact, in view of all the considerations discussed above, we feel that the climatic hypothesis is improbable, that it is inadequate to explain all the facts, and that it is such a weak rival hypothesis to that of earthquake shaking that it cannot be entertained.

Hypothesis of Response to Ablation. Wastage of the lower part of a glacier, as well as addition to the reservoirs must have an effect on glacier variation. It is possible, for example, that the removal of load from a lower part of a glacier might induce a responsive flowage from the heavily loaded part of the glacier above snow line of a nature similar to the flowage resulting from overloading in the reservoir. It is certain that unloading has been in progress in the lower portions of all the advancing glaciers of the Yakutat Bay region.

If the advance were confined to a single glacier, or to a group of glaciers of essentially the same characteristics, this hypothesis might be applicable. But it does not seem applicable to a case in which so many glaciers of such different characteristics are involved. For instance, that underloading in the moraine-covered bulb of Atrevida Glacier should result in an advance in the same years as the advance of Haenke Glacier whose end is in a mountainous valley seems improbable. It is equally difficult to understand by this theory why the Marvine Glacier, which terminates in the piedmont Malaspina Glacier, advanced in the same year as the Atrevida; and also why the other tributaries to the Malaspina have not responded to unloading of the Malaspina by ablation.

Thus, even without considering whether the cause itself is really an adequate possibility for such a spasmodic advance in a single glacier, and without making use of the fact that no such response to unloading is known ever to have occurred, the hypothesis fails and may be discarded in so far as its application to the advancing glaciers of the Yakutat Bay region is concerned.

Hypothesis of Elevation and Tilting. Increase in snow supply to glaciers could be brought about by increase in elevation, and this theory has also been considered. Here again we cannot apply the test of actual measurement, since only the approximate elevations of a few of the peaks are known; but this theory, like that of increase in snowfall, lacks probability. Applying it, for example, to Marvine or Lucia Glaciers, whose sources

are in a lofty mountain region, it is doubtful if an uplift of many hundreds of feet would cause any notable addition to the snowfall. These mountains already rise so high, and form so continuous a barrier, that the damp ocean winds lose a great part of their vapor in rising over them, for the snow line lies between 2000 and 3000 feet. In the case of the smaller glaciers, for example Galiano and Atrevida Glaciers, an uplift might cause distinct increase in snowfall, for the mountains which supply them are lower. But even here, one could not imagine an uplift like the maximum of 47 feet in Yakutat Bay in 1899, or any uplift that was not measured in hundreds of feet, causing a sufficient increase in snowfall to cause such a notable advance and change in glacier condition.

Even making the extravagant supposition of an uplift of several hundred feet, there would not be sufficient increase in snowfall to bring about the spasmodic advance observed here unless with it there was some means of abruptly concentrating the effects of the snowfall on the glacier reservoirs. Such an uplift would also need to be assumed throughout a very great area, for the distance between the sources of the Marvine and the Hidden Glaciers is not less than 60 miles. The extravagant supposition of a mountain uplift of hundreds of feet throughout an area 60 miles in length, even if it were competent to account for the facts, would be proposing as a substitute an even more remarkable phenomenon than that which it is called upon to explain and one not indicated by the delicate seacoast register. Beside the extreme improbability of this hypothesis, even in its least extravagant form, we have the more convincing fact that the advance of the glaciers ceased as abruptly as it began. Were the advance due to uplift, its effects, at least in diminished form, should have continued. For these reasons we cannot entertain this theory.

A sudden advance of glaciers might conceivably be caused by a tilting of their valleys, but this hypothesis certainly cannot be applied to the Yakutat Bay advances, for there are several fatal objections to it in addition to the improbability of a sufficient increase in grade to transform an uncrevassed glacier to broken condition, and to replace long stagnation by sudden advance. The first objection is the lack of evidence by observation. A change in grade sufficient to cause so great a change in glacier condition should certainly be noticeable, and we are unable to detect it. A second objection is the different directions pursued by the different advancing glaciers—some flowing southward, one eastward, and one westward. It would require complexity of breaking and inclination of a mountain region not only entirely without precedent, but, judging by our knowledge of earth changes, wholly impossible. Finally, even if we assumed the possibility of the change it would be necessary to assume also a return to approximately the former condition, for we have to explain not only a sudden advance, but a sudden cessation of advance.

Untenable Earthquake Hypotheses. The occurrence of the severe earthquakes of 1899 naturally led us to consider in what ways it might be possible for the transformation of the glaciers to be a response to these earthquakes, and four hypotheses have suggested themselves, three of which are easily disproved, while one is so supported by the facts that it is retained.

Of the three earthquake hypotheses that are discarded the first is that the changes in the glaciers are the result of breaking of the ice by the vigorous shaking either in 1899 or in some subsequent year. It strains the imagination to conceive of such a shaking as to break a glacier into impassable condition by such a cause, but waiving this, and

assuming its possibility, it is easily proved not to be the case here. This theory fails to account for the advance of the glacier fronts, the spreading at the sides, and the thickening of the glaciers. It fails to account for the breaking of some glaciers and the lack of breaking in others, for the breaking of the Marvine lobe of the Malaspina and the immunity of the rest of this piedmont ice area, and for the appearance of breaking in different glaciers in successive years. It cannot be seriously entertained.

The second earthquake hypothesis is that the earthquakes of 1899 broke the glaciers, but that the effects of the breaking did not appear until ablation revealed the cracks. This theory does not explain the glacier advance, spreading and thickening, and is, consequently, inapplicable here. It is therefore hardly necessary to consider whether a glacier could be broken below the surface and not on the surface, and also whether ablation could reveal the breaking in a period of about 10 months, most of which was winter season, and at the same time in a moraine-covered piedmont bulb as in the clear ice area of the valley glacier as far up as the snow line.

A third earthquake hypothesis is that the severe shaking in 1899 resulted in the setting up of unusual conditions of strain in the ice, or of disturbing equilibrium in the glacier, whose manifold points of contact with the irregular valley bottom and side are postulated to have complex strain relationships, resulting in advance of the glacier in the restoration of equilibrium. No such strain relationships are known to exist and, if they do, it is not known that they are competent to cause forward movement. The hypothesis does not account for the immunity of most of the steep-grade glaciers from advance, for these should seek equilibrium first. It accounts for no supply to keep the forward movement going and thrust a glacier front forward two miles. It may be objected to first on the ground of postulating conditions not known to exist, and, second, in crediting them with great powers where slight power, if any, is more probable.

These three earthquake theories are so entirely inapplicable that it would not be deemed worth while to give space to a consideration of them, were it not for the fact that each has been proposed by a colleague as a possible alternate hypothesis for the theory which is stated in the following section.

Theory of Earthquake Avalanche Supply. This theory is based upon three postulates, each of which is deduced from actual observation. The first postulate is that in normal conditions the mountain slopes, above a certain level, are deeply covered with snow and ice wherever it can stand, and that much of it is in such unstable position that sufficient quantities are normally sliding into the valleys to form the glaciers and keep them supplied. The second postulate is that by September, in any year, a considerable addition to the snow accumulation has been made by the early autumn snows, and that at that season the snow line has been materially lowered by the accumulation of snow, on slopes from which much of it would normally be removed by the next summer's melting. The third and vitally important postulate is that the continued shaking of the mountains during the earthquakes in the first three weeks of September, 1899, dislodged vast quantities of this snow and ice and threw it in avalanches down upon the glaciers and upon the névés. This added to the glaciers a sudden and great supply of snow and ice, some of which never would have reached the glaciers, but most of which would have come down to them from time to time during a series of years.

From these basal postulates we have inferred that so great was the sudden increase of snow and ice in the glacier reservoirs that a wave of advance was started of far greater

vigor than variations in precipitation could cause, and sufficient to explain the spasmodic advance and breaking of the glaciers that has been observed in the Yakutat Bay region. Since the unusual supply ceased as abruptly as it came, the advance quickly ran its course and the glaciers abruptly resumed their former condition.

A number of facts and considerations support this theory, as follows: (1) There was a series of unusually vigorous earthquakes in September, 1899. (2) Accompanying these shocks there was much avalanching. These two points are important in showing that the cause was actually present, though they do not prove that it was adequate. (3) Both the earthquakes of 1899 and the phenomena of spasmodically-advancing glaciers are centered in the Yakutat Bay region. (4) Of the advancing glaciers all but one are known to have begun their advance since 1899 and the evidence is all but conclusive that the other advance, that of the Galiano, occurred after 1899. (5) There has been a rough progression in the glacier advance from the smaller to the larger glaciers, beginning almost at once in the smallest, then affecting larger ones, then still larger, while the largest have not yet responded. (6) The duration of the advance was brief, indicating an intense cause of short duration. (7) The cessation of the advance was abrupt, indicating that the cause terminated abruptly. (8) The extent of the advance was great, involving the transfer of large masses of ice in thickening, spreading and pushing the glacier fronts forward, proving that the cause involved an extensive addition to the glacier supply. (9) The transformation of the glaciers is without recorded parallel, which indicates the necessity of seeking an unusual cause. (10) All other theories fail to satisfactorily account for the phenomena. (11) No facts observed are opposed to the theory of earthquake shaking. (12) The theory satisfactorily accounts for all the phenomena of the advancing glaciers.

It accounts for the location of the advancing glaciers in the Yakutat Bay region; for the beginning of the advance immediately after 1899 in the small glaciers; for its appearance in successive years in other, larger glaciers; for the failure of the largest glaciers to have responded as yet; for the failure of response of glaciers that did not receive numerous avalanches; for the pronounced breaking, advance and thickening of the glaciers; and for the sudden beginning and equally abrupt termination of the advance. Being the only theory that does not fail in some significant respects, having no vital points of objection, but on the contrary being supported by numerous important facts and considerations, and being competent to account for all the phenomena of the advancing glaciers, the theory of earthquake shaking occupies a strong position, falling little, if any, short of complete demonstration.

Nature of the Advance. It is made clear in the preceding chapters that several stagnant or nearly stagnant glaciers, with little if any crevassing, have undergone an abrupt change, becoming impassably crevassed and at the same time becoming thickened at their lower ends and pushing forward at their fronts and margins; and that this period was of brief duration and terminated as abruptly as it began. Such an advance contrasts strikingly with normal motion of glaciers, as well as with the slow advance of a glacier under the influence of normal climatic variations, when through a period of years, and without notable change in surface conditions, the ice front gradually advances.

The reports of the International Committee on Glaciers show that between 1895 and 1907 the overwhelming majority of advancing glaciers in the Alps, Scandinavia, Russia, Canada, etc., moved forward from ten to sixty feet a year, there being one case of an

advance of 114 feet, one of 133 feet, and the remarkable case of the Vernagt-Ferner whose advance attained as many as 150, 240, and 450 feet in a year. In contrast with this the Haenke Glacier in Alaska advanced over 4000 feet in a little less than ten months and the Hidden Glacier over 10,000 feet in a brief interval, probably less than a year and possibly in only a few months. The glaciers in Europe sometimes continue to advance for fifteen or twenty years when once an advance starts, the cycle from maximum to minimum being 35 to 50 years, contrasting with the Alaskan glaciers under discussion whose advances have lasted less than a year.

The case of glacier advance most nearly analogous to that of the Yakutat Bay region, of which we have found record, is that of the two small Tyrolese glaciers—the Vernagt and Ferner, which are known to have undergone great advance in 1599, 1680, 1773, 1820, 1845 and 1898–1900. Some of these advances have been very great, and by their effect in damming up a tributary valley and forming a lake, whose sudden drainage has caused great floods in a populated valley have attracted attention and invited study even in early times. Reference to these earlier studies will be found in the remarkable monograph by Finsterwalder[1]; in which the cause for the variation in the Vernagt and Ferner Glaciers is discussed. The notable advance of the Vernagt Glacier and the less marked advance of the Ferner Glacier in 1898–1900 are described by Blümcke and Hess; and they have also published some of the results of careful studies of these glaciers upon which they have been engaged for many years.[2]

In the last advance of the Vernagt glacier a swelling was at first observed in the reservoir, and later a transformation of the ice tongue followed, converting it from practically uncrevassed to crevassed condition, and causing an advance of the end. Earlier advances had been similar, though some of them were far more pronounced. There is therefore a noteworthy resemblance between the behavior of the Vernagt-Ferner and the advancing Yakutat Bay glaciers, though the 1898–1900 advance was much less spectacular and less rapid than those of the Yakutat Bay glaciers, and there is also a difference in the cause, for the Vernagt, and less notably the Ferner, is a small ice tongue fed from a very large reservoir, and the advance is ascribed to the concentrated influence of an increase in the snowfall in the reservoir, not to the effect of earthquake shaking.

From the researches of Finsterwalder, Blümcke and Hess it seems evident that excessive accumulation of snow in the reservoir ultimately starts a wave which, gathering force, finally sweeps with rapidity through the glaciers and finds response even in the lower portions of the glaciers by rapid motion, thickening, and advance. Their important detailed work, which, in the nature of the case, could not possibly have been undertaken in the Alaskan glaciers, throws a flood of light upon the problems of ad-

[1] Finsterwalder, S., Der Vernagt-Ferner, Wissenschaftliche Ergänzungshefte zur Zeitschrift des D. u. Ö. Alpenvereins, I Band. 1 Heft. Graz, 1897; with addendum by Blümcke, A. and Hess, H. on Die Nachmessungen am Vernagt-Ferner in den Jahren 1891, 1893 and 1895.

[2] Blümcke, A. and Hess, H., Beobachtungen an den Gletschern des Rosenthales. Mittheilungen des D. u. Ö. Alpenvereins, 1900, Nr. 4, 6 pp.; Einiges über den Vernagt-Ferner, ibid. 1902, Nr. 18; Tiefbohrungen auf dem Hinterreisferner, 1902, ibid. Nr. 21, 7 pp.; Tiefbohrungen am Hintereisferner im Sommer, 1908, Zeitschrift für Gletscherkunde, Band III, 1909, pp. 232–236; Tiefbohrungen an Hintereisgletscher, 1909, ibid. Band IV, 1909, pp. 66–70; Hess, H. Zur Mechanik der Gletschervorstösse, Petermanns Geogr. Mitteilungen, 1902, Heft. V; Probleme der Gletscherkunde, Zeitschrift für Gletscherkunde, Band I, 1906, pp. 241-254; Die Gletscher, Braunschweig, 1904. pp. 296, etc.

vance of these glaciers, and in formulating a theory of the nature of the advance we naturally start with the results of their work as a basis. The advance in the two regions is closely analogous in main characteristics though differing in cause, rate of transformation, and amount of forward motion; but the two latter differences may well be no more than what would normally result from the difference in cause,—in the one case the slow accumulation of energy and ultimate response from increase in snowfall, in the other, rapid application of the same cause due to sudden precipitation of snow into the reservoirs from the valley walls and corresponding rapid and great response in the lower glacier.

In explaining such an advance of glaciers as those described in this and preceding chapters we cannot assume actual flowage or transfer of ice from reservoir to terminus because of the brevity of the time and the abruptness with which the advance passed through the glaciers. It is inconceivable, for instance, that there was a transfer of material through the 15 miles of the broken Marvine Glacier, in the brief period of a year or less. Snow that falls in glacier reservoirs, changed to ice, does by normal glacier motion ultimately, in the course of a score of years, find its way to the glacier front, but we imagine that no one would advance the hypothesis that glacier motion could under any circumstances become so accelerated as to permit such transfer as would be called for by the Alaskan glacier advances.

A modification of this hypothesis that is less difficult to conceive is that, by increase of snow in the reservoir, a condition of advance is started in the upper glacier, and that this is so great and powerful that it results in a rapid actual forward transfer of ice, causing motion in the glacier of far greater rapidity than normal. By this rapid advance the glacier surface is broken and a thrust is applied against the ice in front which is not yet in such rapid motion. This thrust pushes the ice forward and thus extends the breaking and advance beyond the zone actually reached by the ice which is moving rapidly forward. In other words, this hypothesis takes as the basis the normal means of transfer of ice in glaciers, but supposes an acceleration in speed, and adds to it the pushing forward and consequent breaking up of the rigid stagnant outer ice, much as an advancing glacier would push forward the wall of a building that stood in the way.

In 1906 this hypothesis seemed, in part at least, a possible explanation of the conditions in the advancing glaciers of Yakutat Bay. In the light of the observations of 1909, and after further consideration of the problem, it now seems to us an inadequate explanation, for there are several serious objections to it. In the first place, it is doubtful if, even with the most rapid glacier motion we could assume, the wave of advance could have extended as far down the glacier by 1906 as this explanation would demand; and, in any event, it should have been preceded by breaking in the upper glacier, and not have abruptly affected the entire glacier, for many miles, all in a period of a few months. Furthermore, it seems exceedingly doubtful whether sufficient thrust could be applied to so completely break the surface of such extensive areas as were broken, for instance in the Marvine lobe of the Malaspina by the 1906 advance. Even more difficult to conceive is the application of a thrust of sufficient force to push the front of Hidden Glacier forward two miles. Finally, the fact that the advance ceased abruptly and completely immediately after the breaking of the ice was accomplished is a serious objection. Had there been an active advance from behind, pushing

the stagnant ice forward and breaking it, there could hardly have been so abrupt a termination, for after the first thrust further flowage should cause continued advance. Because of these objections we feel forced to set this hypothesis aside as improbable.

Another conceivable hypothesis is that the sudden addition to the upper portions of the glaciers started a wave in the ice, which was transmitted down-stream, and which was transmitted wholly independently of glacier motion. As this wave paused down the glacier it broke the ice surface, and since it moved with rapidity it transformed the glacier from end to end in a brief period of time; and after it had passed, the ice returned to its normal state and the broken surface was given over to the processes of ablation to heal. We confess that we find this hypothesis difficult of conception, and that the form and nature of movement of the wave is not clear. If, however, we could grant the passage of such a wave, it would explain some of the phenomena of the advancing Yakutat Bay glaciers, though there are many which are not capable of explanation by such a wave. This hypothesis, for instance, would not explain the notable advance of the front of the glaciers. To carry an ice front forward one mile, as in the case of Haenke Glacier, and two miles, as in the case of the Hidden Glacier, demands an actual transfer of ice. With such a passage of a wave through the glacier, so far as we are able to conceive it, we should expect that when it reached a stagnant piedmont bulb it would spread and gradually die out, as was the case in the Variegated and Atrevida piedmont areas. But in Malaspina Glacier the breaking extended many miles down the Marvine lobe, without spreading laterally westward; in other words, it followed the flow lines influenced by the Marvine. A wave in the ice might be expected to disregard this factor and pass through ice regardless of its source of origin; and a wave derived from an impulse set up in the Marvine Glacier valley, would, it seems to us, on passing out into the Malaspina Glacier, as easily spread out westward into the ice supplied by the Seward Glacier as southward along the Marvine lobe. A third objection is the thickening of the ice, especially noticeable in the piedmont areas of the Variegated Glacier and the western margin of the Atrevida Glacier. In these places the ice has the appearance of having flowed out there as a thick viscous fluid flows. In this connection may be mentioned the development of the clear ice area in the Atrevida Glacier which we are wholly unable to explain on the hypothesis of the passage of a wave through the glacier. These facts make it evident that a mere wave is inadequate as an explanation of the phenomena of the advancing Yakutat Bay glaciers. Any theory to account for these phenomena must explain a rapid transfer of impulse from far up the glaciers, probably from the reservoir, through many miles of hitherto stagnant or slowly moving ice, and accompanying this a breaking of the glacier surface and an advance of the front, a spreading of the margins and a thickening of the glacier. A combination of the theory of wave motion and actual flowage will account for these phenomena, though neither one will do so by itself, nor, for that matter, will any other theory of which we are cognizant. We are therefore proposing, as an explanation of the behaviour of the Alaskan glaciers an hypothesis which we call the *glacier flood hypothesis*.

Under the glacier flood hypothesis it is assumed that the ice of a glacier is rigid in its upper portion and along its margins, but that it is in a state of viscosity beneath a rigid crust; that is, a glacier has its zones of fracture and of flowage, like the earth's crust. It is further assumed that the degree of viscosity is increased by pressure, and

that with the development of viscosity the rate of possible ice flow is increased. The maximum thickness of glaciers is rather small for great increase of viscosity by overloading the upper layers, but the pressure inferred would be not only direct vertical pressure of upper layers on lower, but also hydrostatic pressure of heavily-loaded upper portions of the glacier stream on the parts of the ice farther down-stream and down grade from the portion overloaded by avalanches. In narrow glaciers these avalanches doubtless went clear across the ice tongue, and where they did not do so, the same marginal area was overloaded. The final assumption is that the addition of snow to the upper glacier by the effect of earthquake shaking started a wave of advance which by applying pressure to the ice in front induced such a degree of viscosity that flowage followed, and that continued and was transmitted down the glacier until the effects of the advance wore out either because the end of the glacier was reached or because of dissipation in an expanded piedmont bulb.

By the glacier flood hypothesis it is not conceived that there was actual transfer of ice from the glacier head to the glacier end within the brief period occupied by the advance, but that there was, nevertheless, actual viscous flowage of the lower ice throughout practically the entire broken area, the state of increased flowage being progressively extended down the glacier until the effects of the thrust died out. That is to say, there was a wave-like progression of a thrust down the glacier which, when applied, temporarily increased the viscosity of the lower ice layers, and this was followed by a rapid increase in the ice motion. At some time during the period the viscous lower ice layers of practically the entire glacier flowed forward to a greater or less distance, but no single portion of the ice advanced even a considerable part of the distance from the head to the end of the glacier. With the passage of this wave and the accompanying flowage of the ice, the rigid upper crust of the glacier, and the rigid margins, were subjected to such strains that the glacier was broken into the condition observed in the advancing glaciers. The hypothesis further conceives that there was some breaking of the margins, notably in the thinner parts of the piedmont areas, by a rigid thrust without the immediate association of flowage.

This hypothesis has been called the glacier flood hypothesis because of the resemblance to a river flood. When rain falls heavily, or snow melts rapidly, in the upper reaches of a river the effect of the sudden increase in volume extends down-stream, and if the river surface is covered with ice this is broken and piled up as the rising flood passes beneath it. The condition of the advancing Yakutat Bay glaciers is conceived by this flood hypothesis to imitate the river flood in such important ways as to warrant a comparison between the two; but there are two notable differences, first the difference in rapidity of transfer of the flood effect in the two cases, and secondly the difference in nature of the flowage. In the river the flow involves the transfer of water from the flood source to the river mouth; in the glacier there is no such transfer. There are other differences of minor character, but of scarcely enough importance for statement here.

The glacier flood hypothesis has the one merit that all the other hypotheses lack; it seems to account for all the facts. With the passage of such a wave and the associated rapidity of flowage of the under-ice it is easy to understand the accompanying breaking of the surface ice. The theory accounts equally well for the fact, observed in the advancing Lucia Glacier, that the margins of the glacier are more broken than

the center, as if there were a dragging and a lateral shove there. It explains the lateral spreading of the glaciers and the forward motion of their fronts by the transfer of ice crowded both laterally and forward by the flowage of the lower ice under the pressure of the thrust from behind. It accounts for the uprising of clear ice areas in the Atrevida and Variegated Glacier bulbs, alluded to on a later page, and for the notable thickening of portions of these bulbs. By this hypothesis it is easy to understand why the breaking of Malaspina Glacier did not extend far westward into the Seward lobe, but was practically confined to that portion of the Malaspina whose ice supply is furnished by the glacier which gave rise to the wave of advance. Equally well does the glacier flood hypothesis explain the abrupt beginning and abrupt ending of the transformation, for the advance would not begin until the thrust was applied, and its effects would terminate soon after the wave had swept on.

Since the glacier flood hypothesis explains all the facts, while other hypotheses fail in significant respects, we believe that it occupies a strong position. There are, however, two objections which will appeal more or less strongly to those who give this hypothesis consideration. One of these is the rapidity with which the breaking extended throughout the glaciers. It is to be noted, however, that this objection would be equally strong against either of the other theories. The rapidity of progress of the advance and breaking, when once begun, is a fact fully established by observation, and whatever theory is proposed in explanation of the change must of necessity assume that the cause was capable of operating with the rapidity observed. So far as we can see it is fully as easy to conceive such a rapid transformation under the glacier flood hypothesis as under any of the other hypotheses. In the absence of more exact knowledge of the behavior of ice in glacier motion, it is not possible to give to this problem the thorough consideration which it doubtless deserves.

If we may assume for glaciers a state of viscous flowage in those parts of the ice under sufficient pressure,—which some will deny,—it seems reasonable to infer that there will be an increase in liquidity under increase in pressure. Our hypothesis is that the application of pressure from the thrust did thus increase the liquidity of the ice, and to a sufficient degree to bring about the phenomena of advance, thickening and breaking observed. This hypothesis is based upon the necessity of such a result from the phenomena observed in the field, which force us to the theory of viscous flowage as postulated under the glacier flood hypothesis. We make no pretention of discussing the physical aspect of the subject, and until there is more knowledge and fuller agreement upon the simplest principles of the behavior of ice in moving glaciers we feel that there would be little profit in such a discussion.

The second objection which some will urge to the glacier flood hypothesis is that already hinted at in the last paragraph, namely the doubt whether glacier ice does under any conditions behave as a viscous body. Upon this subject opinion is divided, and while some believe in viscosity of glacier ice under pressure, others are equally convinced of its impossibility. Much has been written, on this subject both by physicists and by geologists, and since the result of this discussion has been only to lead to a divided opinion, and since the discussions are accessible to all, and known to most students of glacial phenomena, it does not seem necessary to restate or even abstract them here. There are, however, some phenomena of direct observation in the advancing glaciers of Alaska, and some considerations deduced from these observations, which

seem to us to have such direct bearing on the problem of glacier motion, and to point to viscous flowage as the mode of motion, that we deem them worthy of specific statement in this connection.

Evidence of Viscosity. If it is granted that there has been actual transfer of material in the advancing glaciers of Yakutat Bay, and the facts render no other conclusion possible, we may ask whether any of the other theories of glacial motion than that of viscous flowage are capable of accounting for the abrupt transfer of ice to such a degree and extent as is observed? As a result of our consideration of this question we have been forced to the conclusion that no other mode of motion than viscous flowage can be appealed to in explanation of the thickening, spreading, and advance of these glaciers. We are, of course, open to conviction of error in this conclusion if it can be shown by those who accept other theories of glacial motion that any one of them can explain the phenomena of the advancing glaciers. The motion, whatever its nature may have been, must explain the faulting and linear crevasses on the Lucia Glacier, the increase of breaking along the margin and the piling up of ice on the stoss side of the Lucia nunatak. It must explain the system of crescentic crevasses in the Atrevida Glacier piedmont bulb, and the pronounced thickening and spread of this bulb and that of the Variegated Glacier. Further, it must account for the rapidity of all the changes and the notable forward motion of the fronts of Haenke and Hidden Glaciers, one and two miles respectively. All these phenomena are but natural and simple results of viscous flowage if once the possibility of this is granted; are they also capable of explanation on the basis of other theories of glacier motion? We cannot conceive that they are and, consequently, because of the failure of other hypotheses are driven to believe in viscous flowage as the mode of motion in these advancing glaciers.

In addition to this consideration, there are some significant facts bearing on the problem of the mode of motion of the glaciers. One of these is the appearance of clear ice areas in parts of the glacier surface which prior to the advance were overspread with ablation moraine. This is observed on the Variegated Glacier bulb where previously-existing moraines were swept forward, the area of clear ice extended beyond the mountain valley to a portion of the surface formerly covered by moraine, and a new area of clear ice introduced between moraines far out in the area of former ablation moraines. It is even more clearly illustrated in Atrevida Glacier, where an extensive area of clear ice appeared in 1906 in the midst of moraine-covered ice and in a position where in 1905 there was a uniform sheet of ablation moraine. Here also the ice layers are highly inclined as if there was uprising from below. We can readily explain these facts on the hypothesis of viscous under-flowage and local uprising of ice from below, pushing the moraine-covered ice aside in certain favorable localities, the ice becoming crevassed after it rose to the surface and was released from strain, as 1906 photographs show. We are wholly at a loss to account for it on any other mode of ice motion.

That the lower ice of a glacier is in a state resembling plasticity, and wholly in contrast to the upper ice, is suggested by the crevasses which break so many glacier surfaces. But in the advancing glaciers of Yakutat Bay we have a better chance of studying and understanding these crevasses than usual. They did not exist before the advance, they were then developed in extraordinary numbers, and then at once their cause ceased and they began to be destroyed by the ablation of the ice surface. The conditions under which these crevasses were developed, we may fairly assume, were

such as to give rise to the deepest possible breaking. We, therefore, have observational data for tracing the life history of a system of profound crevasses; and the result is, what is well known from the study of other crevasses more normally produced,—namely, that they are mere surface fissures in the very upper parts of the glacier. One or two hundred feet of ablation lowers the ice surface to their bottoms, and below that the ice is unbroken. Yet surely the lower ice was subjected to a profound straining of such abrupt nature that it should have responded by fissuring if its condition had permitted. Ice in a state of plasticity such as the theory of viscous flowage demands, could not be fissured by even the abrupt and powerful strains which accompanied the advance of the glaciers; but ice in a condition demanded by theories not involving a state of plasticity would be expected to yield to such strains by fissuring.

We cannot observe the condition of glacier ice in motion, and experimentation has so far not even approximately duplicated the conditions existing in the interior and bottoms of glaciers. These facts account in part for the failure of inferences from theoretical considerations of the problem to bring forth results which are convincing to all. In these studies of the problem, known facts have been considered in the development of theory, but the facts are partly unknown, and there is necessity of assumption which if incorrect removes the foundation of the theory. In all of the theories of glacial motion there are assumptions upon which there is not general agreement,—hence the failure of even the most elaborately argued theories to receive universal acceptance. The theory of viscous flowage has lost favor with some students of the subject of glacial motion by reason of the fact that glacier ice is a crystalline rock; and the assumption has been made that this condition and the supposed inability to hold stones in the bottom and perform glacial erosion are incompatible with viscosity. Partly upon the basis of these assumptions several elaborate hypotheses have been put forth as a substitute for the theory of viscous flowage, which has numerous adherents.

Still maintaining our determination not to enter into a discussion of the literature on the physics of glacier motion, and confining ourselves primarily to a statement of facts of observation and the conclusions to which they lead, we will state some facts observed in the Yakutat Bay region which we believe have a direct bearing on the problem of the condition of glacier ice. While there, as in other regions, we cannot make direct observations on the condition of a large glacier from top to bottom, we nevertheless have samples of the glacier ice, from different parts of the glacier, supplied to us for observation in the form of innumerable icebergs, floating in the waters of the fiord and stranded on its beaches. From a study of these and of the clear glacier surfaces, sides of crevasses, and ice cliffs of the glaciers we can form some idea of the nature of the glacier from top to bottom.

The largest of the glaciers, the Hubbard, presents an ice cliff of marble whiteness some three hundred feet high, the upper half more or less profoundly crevassed. What lies beneath the water level is hidden, but the appearance of icebergs black with débris, and the abundance of débris-charged bergs in the waters of the fiord lead us to conclude that the very lowest layers are heavily burdened with morainic materials. Among the thousands of icebergs floating in the fiord and stranded on its shores about two-thirds are whitish and one-third are glassy, as shown by repeated observations at favorable localities (Pl. LXXIX, B). Some large bergs are entirely glassy, some entirely whitish, and some have bands of white and glassy ice, the proportions varying in dif-

ferent bergs. The white bergs owe their color to the presence of great numbers of included air bubbles, all of small size, and often very closely spaced; the glassy bergs have fewer air bubbles and sometimes none at all. In general the white bergs are the more finely-crystalline, though many of them are coarsely-crystalline; but the glassy bergs are always coarsely-crystalline. We assume that the air bubbles of the white bergs represent air imprisoned in the course of snow accumulation in the upper portions of the glacier and included in the ice during its crystallization.

Why is there this difference in the ice from the same glaciers? The obvious answer is that it is due to position in the glacier, the glassy, coarsely-crystalline bergs being from the lower ice, the more finely crystalline, air-charged, white ice from the upper portion; but in those frontal portions of glaciers where ablation has removed much of the upper ice, coarsely-crystalline ice forms the surface. This conclusion is supported by two facts: (1) the part of the glacier front above the water in the ice cliffs is white; (2) although some white bergs contain débris, the most heavily charged icebergs are invariably glassy. The problem then resolves itself into the question, why is the upper ice charged with air bubbles while in the lower ice there is no such abundance of air bubbles? We have tried to conceive of conditions under which this difference in ice condition could be brought about. It is inconceivable that the lower ice originally contained less imprisoned air than the upper ice, for it is from the same source of snowfall, compacted and consolidated into the crystalline ice. We are forced to conclude, therefore, that the air originally included in the lower region of glassy ice has been eliminated in the course of development of the coarsely-crystalline structure. That such elimination has been in progress within the glacier is indicated by the fact that air bubbles are present in great abundance in some coarsely-crystalline ice and in all the finely-crystalline bergs. By this fact we are forced to the conclusion that in some parts of the ice there are unusual conditions which permit the elimination of included air bubbles and that these conditions prevail mainly, if not entirely, in the bottom layers of the glacier.

From the consideration of the problem which we have been able to give we can find but one explanation that is capable of accounting for these facts. We accordingly propose as an hypothesis that the lower portion of the glacier is in a condition of sufficient viscosity to force the escape of the included air bubbles, which, rising, become included in the upper area of viscosity, giving rise to that layer of ice from which some of the whitest icebergs are derived. Whether the coarsely-crystalline structure is present in the viscous ice, or whether it develops as the front of the glacier is neared, and the pressure relieved, is not indicated by any facts which we have observed.

From the above statement of facts, and from the consideration of the inferences which seem to be warranted from them, we are led from various points of view to the same result. We are wholly unable to understand the phenomena observed in the Yakutat Bay region without resort to the theory of viscous flowage. By this theory all phenomena are readily explained, while by any other theory we find ourselves confronted by grave difficulties in accounting for some of the phenomena. We are, therefore, forced to believe that the theory of viscous ice motion has foundation, and that the phenomena of the advancing glaciers of Yakutat Bay are an expression of an unusually active phase of such flowage.

Lack of Similar Advance Elsewhere. Were it not for the spectacular nature of the

transformation of the advancing Yakutat Bay glaciers it might be urged against the theory of advance under the impulse of earthquake shaking that the fact that analogous advance in other regions had not been observed was opposed to this theory. Such an objection would have some force if the advance were the result of ordinary changes, to be commonly expected in regions of glaciation; but it lacks force in view of the exceptional conditions under which such an advance is possible. Only by the combination of vigorous earth shaking and heavy snow cover in unstable positions, can such a notable advance be brought about. The latter condition is commonly present in the higher mountains from which good-sized glaciers flow, though rarely in such a degree as in the Alaskan coast ranges; but vigorous earth shaking is not so common. Even in Alaska, a region of abundant earthquakes, such shaking as that of 1899 must be rare; and although we have record of many earthquakes, both before and after 1899, we have heard of no shaking in Alaska between 1788 and 1913 that equals this period either in intensity or in duration. The Alps, whose glaciers have been studied with greater care and for a longer time than any other region, is not a section of abundant, great earthquakes. In other regions of earthquake-shaken mountains, like the Caucasus and the Himalayas, the glaciers have been little studied. We hold, therefore, that in some glacial regions an advance even approximating that of the Yakutat Bay glaciers is not to be expected; and even in those regions where a similar transformation may be looked for, it could come only at intervals many years apart. It is our opinion, that the phenomenon, in anything like the degree observed in the Yakutat Bay region, is of rare occurrence, which is the main reason why it has not hitherto been observed.

There is the further fact that the cycle of the advance is completed in such a brief period of time that an observer must be on hand at just the right time to detect it, and he must have knowledge of the previous condition of the glacier and the position of its front. For instance, if we had not had Russell's description and photographs of 1890 and 1891 we would not have known of the change in Galiano Glacier; had Gilbert not observed Variegated Glacier in 1899 and we in 1905, we would in 1906 have naturally assumed that the broken condition was normal to it. Similar statements might be made about the other advancing glaciers. Moreover, even if we had previous observations as a basis for comparison, since the advance stops so abruptly, in a year or less, not going on for several years, as under normal climatic control, and since ablation so quickly heals the broken surface, a cycle of advance might pass completely without detection, if the period between observations were five or ten years. It is also to be noted, that if a glacier surface were normally crevassed, as Hubbard, Turner, and Nunatak Glaciers are, the evidence of change would not be so easy of detection as in the uncrevassed, stagnant glacier ends.

By good fortune our expeditions to Yakutat Bay were timed just right for the detection of the advances, and they had the advantage of having available the results of the previous expeditions by Russell and Gilbert. There may have been similar advances in other remote regions of heavy snowfall and great earthquakes, say in Alaska, or the Caucasus, or the Himalayas, which have failed of detection because of the lack of successive expeditions properly timed to note the evidences of change. Hence, not only are there few regions where spasmodic advance of glaciers in response to earthquake shaking is to be expected, and even there only at long-separated intervals; but the chance of discovery of the evidence of such advance, and the correlation of it with

its cause, in those remote regions where the conditions are favorable, but which are as yet little studied, is so slight that, even though advance occurs, it may be undetected.

Other Advancing Glaciers. With the discovery of a new cause for the advance of glaciers, which cannot have been considered in the explanations of glacier advances observed prior to this, we naturally consider the question whether the new theory will not explain some of the old advances. We have not found time to make an exhaustive study of the literature of advancing glaciers outside of Alaska to see whether there are indications that this new theory will apply to any of the old cases. We have been deterred from undertaking it partly because of the desire to have the theory discussed and further observations made first, and partly because there seems at least an even chance that such a study of the literature would end only in inconclusive results, since critical observations may not be recorded. It is, for example, possible that the reported great advance of some of the Himalayan glaciers, and the notable advance and transformation of Sefström Glacier in Spitzbergen between 1882 and 1896, and of the neighboring Wahlenberg Glacier between 1896 and 1908, while other neighboring glaciers were receding, may be a response to the influence of earthquake shaking of an undetermined period.

There are two points of a general nature bearing on this subject which we wish to make before considering the possible extension of the earthquake theory to other advancing glaciers of Alaska. One of these is the proposition that in small glaciers with broad reservoirs, like the Vernagt Glacier in the Tyrol, the phenomena of advance as the result of the fall of unusual numbers of avalanches without earthquake shaking would imitate those of the advance of larger glaciers under the impulse of more avalanching under vigorous earthquake shaking. Two cases of the latter sort are the little Boveyre Glacier in Valais, Switzerland, which advanced because of great increase of material due to an avalanche, and the Stutfield Glacier of Canada which advanced about half a mile and destroyed a fringing forest, when covered with débris by great avalanches.

The second proposition is that in larger glaciers moderate avalanching under the influence of moderate earthquake shaking would result in a response of far less vigor and extent than that observed in the Yakutat Bay glaciers. If this proposition is correct, it is reasonable to expect that some advances of moderate amount, and even without accompanying breaking of the surface, may follow earthquake shaking in the snowfield region. It is well established that there are moderate fluctuations of glacier fronts in the Alps, and elsewhere, as a result of climatic variations; but there are also exceptional changes in position of the ice fronts. The application of the avalanche theory, either as a result of earthquake shaking or of other cause, to some such variations in glaciers is possible. It will be interesting to see the theory put to a test in some such case.

In Alaska, outside the Yakutat Bay region, there have been many changes in the position of glacier fronts within the period of observation, and some of these may be due to the effects of earthquake shaking. Unfortunately observations on most of these glaciers are of too limited a nature, and for too short a time, and our knowledge of the extent and nature of preceding earthquake shocks, and of the amount and variation of the snowfall of the Alaskan coastal region, is too slight to permit us to enter into a full or conclusive discussion of the possible cause for these variations. Yet there are some notable facts leading to tentative conclusions, which may be stated.

In 1892 Russell wrote his paper on Climatic Changes Indicated by the Glaciers of

North America,[1] in which he states that "the evidence that a general retreat of the glaciers of Alaska is still in progress is abundant, and in a few instances of quantitative value." In Alaska the only advances mentioned by Russell were the comparatively recent advance of part of the border of Malaspina Glacier near Pt. Manby, from which the ice had retreated 1500 feet in 1891 when observed by Russell, and the advancing Frederika Glacier of the Wrangell Mountains, observed by Hayes in 1891, and cited by Russell as "the only instance of an advancing glacier known on the west coast of North America." He concludes that "the data relating to both the fluctuations of glaciers and to climatic changes are inadequate for satisfactory comparison," and that "the growth of glaciers and the initiation and decline of glacial epochs, are caused by very gradual climatic changes which would only become conspicuous, as climatic changes are now studied, after the lapse of centuries."

Gilbert, after his glacier studies in 1899,[2] when he mentions a few other advances and retreats not cited by Russell, concluded that the lack of parallelism between variations of glaciers and different groups of glaciers in Alaska whose disparities are of larger scale than those of the Alps, just as the glaciation is of larger scale, might not be explained by either of two prominent climatic hypotheses, to each of which there are difficulties of application because the glacial and climatic data are indefinite, and because in climatic units in Alaska there is no known unity of glacier variation. He felt, however, that the explanation must be climatic rather than diastrophic and made a suggestion that "the combination of a climatic change of a general character with local conditions of varied character, may result in local glacier variations which are not only unequal but opposite." The climatic change postulated is a change in the temperature of the water of the Gulf of Alaska. With the water becoming warmer and all other factors affecting glaciation remaining unchanged, "the consequences would include:

1. A higher temperature for the air currents flowing from the gulf to the land.
2. A greater contrast in temperature between the coastal belt and the interior of Alaska, especially in winter.
3. Greater evaporation from the ocean and a higher humidity for the landward-flowing air—resulting from 1.
4. Greater precipitation on the mountains, especially in winter—resulting from 2 and 3.
5. A shorter annual period in which precipitation takes the form of snow—resulting from 1.
6. A (probably) lower ratio of snow to rain—resulting from 5, qualified by 4.
7. A higher snow line.
8. More rapid waste of ice and snow by evaporation and melting—resulting from 1, 5 and 7.

Of these consequences, the increase of precipitation would tend to enlarge glaciers, while the lessened ratio of snow precipitation and the enhanced wasting would tend to reduce them.

Evidently a lowering of the temperature of the gulf water would be followed by the reverse consequences."

As is shown in other chapters of this book, there have been many other glacial advances

[1] Russell, I. C., Amer. Geol., Vol. IX, 1892, pp. 322–336; Glaciers of North America, Boston, 1896, pp. 146–159.

[2] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 102–112.

in Alaska, besides those in Yakutat Bay.[1] For example, Muir Glacier advanced about 300 yards between 1890 and 1892; Patterson Glacier was advancing and destroying trees in 1891; Taku Glacier had a slight net advance between 1890 and 1905; Brady Glacier advanced 5 miles between 1794 and 1894, the destruction of trees being in progress in 1880; LaPerouse Glacier was advancing and destroying forest in 1895, but ceased before 1899, readvancing into the forest once more between 1909 and 1910; two of the Lituya Bay glaciers advanced two and a half and three miles respectively between 1786 and 1894, one advancing another half mile between 1894 and 1906; the Alsek river terminus of the Grand Pacific through-glacier was advancing and destroying forest in 1908; another glacier coming into the Alsek valley from the west, and connected with the Hidden-Nunatak-Yakutat glacier system, though inactive in 1906 and 1908, possibly advanced in 1909; a small glacier opposite the Frederika Glacier in the Wrangell Mountains was advancing in 1908; Childs Glacier advanced in 1910; Logan Glacier, north of Mt. St. Elias, advanced in 1912, after at least 200 years of inactivity; Rainy Hollow Glacier advanced 2000 feet in 3 months in 1910; Rendu Glacier in Glacier Bay advanced 1½ miles between 1907 and 1911; an adjacent cascading glacier advanced ½ mile in 1911; Grand Pacific Glacier advanced about 4000 feet between August, 1912, and September, 1913, and three other ice tongues in Glacier Bay were moving forward in 1913. In these and many other parts of Alaska some glaciers have, in general, been retreating steadily within historic times.

Our own view would now be that any or all of the advances of Alaskan glaciers cited might be due to earthquake avalanching rather than climatic variations, the earthquake data being no more incomplete than the climatic data. Indeed, in view of the Yakutat Bay cases, the greater advances, such as the 20 mile advance of Malaspina Glacier described in Chapter III, the 5 mile movement of Brady Glacier and the 2 to 3 mile advances of the Lituya Bay glaciers, are far more suggestive of spasmodic earthquake advances than of climatic variation, and some of the smaller ones may be due to earthquakes and some to climatic oscillations. These glaciated mountains are known to be frequently shaken by severe earthquakes.[2]

If avalanching during the 1899 earthquakes of Yakutat Bay caused the advance of the glaciers there during the ensuing ten or more years, then it ought to be carefully considered whether similar avalanching in the snow-laden Fairweather Range might not have caused the 2 and 3 mile advances of the Lituya Bay glaciers between 1786 and 1894 as a result of the earthquakes felt at Sitka in either 1843, 1847, 1861, or 1880; whether the same earthquake did not cause avalanching which resulted in the 5 mile advance of Brady Glacier between 1794 and 1894; and whether the earthquake of 1900 in the Chugach Mountains did not similarly cause the slight advances of Valdez, Shoup, Columbia and adjacent glaciers, as some earlier earthquake may have caused the 1892 advance of Columbia Glacier, the 1890–92 advance of Muir Glacier, and the others cited. Many of the minor oscillations like the last mentioned might be climatic, though that leaves their localization unexplained. It is just as logical to assume that they were not climatic, since the earthquake explanation does account for the localization, as it is to assign to them a climatic cause.

[1] For a more complete list of such advances, see Martin, Lawrence, Guidebook No. 10, Excursion C 8, International Geological Congress, Ottawa, 1913, pp. 157–162.

[2] Martin, Lawrence, The Alaskan Earthquakes of 1899, Bull. Geol. Soc. Amer., Vol. XXI, 1910, pp. 339–406; Tarr, R. S. and Martin, Lawrence, Professional Paper 69, U. S. Geol. Survey, 1912.

Application to Former Greater Expansion of Glaciers. In Yakutat Bay itself there has been a comparatively recent advance of the glaciers, during which Nunatak and Hidden Glaciers pushed forward until they united and extended their combined glacier front far up Russell Fiord.[1] The Russell Fiord Glacier was then between 15 and 20 miles farther down the fiord than the front of Nunatak Glacier has been at any time since its discovery, and between 10 and 15 miles farther than Hidden Glacier front. The expansion of Turner and Hubbard Glaciers may have been less notable, but still great, a branch of the expanded Hubbard-Variegated Glacier perhaps joining the Russell Fiord Glacier. The evidence of this advance is a series of stratified glacial gravels overridden by the ice and thereby sculptured by glacial erosion and veneered by a deposit of unassorted till, bowlders, and other glacial débris, while at the head of Russell Fiord is an abandoned lake beach, formed by the waters held in behind the ice dam. That the period of ice advance was brief is proved by the fact that its erosion did not succeed in removing the gravels accumulated before the advance. Until 1906–07 Hidden Glacier was still receding from its advanced position, and recession was still in progress in Nunatak Glacier till 1909–10. The evidence that the recession had been rapid, and that the period even of maximum expansion was recent is complete, for on the outer part of the area reached by the glacier, and on and below the level of the lake beach, the forest has not yet reached the mature stage, while between this and the glacier the vegetation diminishes in age and stage of development, and for a distance of several miles from the glacier fronts only annual and perennial plants and individual young alder and willow bushes are growing. Similar conditions occur in front of Hubbard Glacier in Russell Fiord and Disenchantment Bay.

It is noteworthy that this advance followed a period during which the glacier fronts had been farther back than now, and which was of such long duration that mature forests grew on the shores of the fiord over which the advancing glaciers swept, and even up the valleys where ice now stands, as is conclusively shown by fragments of wood now being carried out by the glaciers from regions at present not forested.

Northeast of Yakutat Bay, on the other side of the St. Elias Range, are the Dezadeash and Kaskawulsh valleys, within which McConnell[2] found convincing evidence in 1904 of former, long, narrow lakes which he believes to have been dammed by the expanded glaciers of the upper Alsek region, that is by ice tongues connected with the through-glacier systems that reach Yakutat Bay. The preservation of driftwood and the lack of mature forest upon the abandoned lower beaches of these lakes, which were 150 feet deep, and the legends of the natives suggest that the advance took place less than a hundred years ago, probably indicating a glacial expansion here contemporaneous with the last maximum of the Yakutat Bay region. The youth of the trees growing on these lower beaches, the spruces not being over 3 inches in diameter, proves how recently this advance took place.

Brooks recognized the beach terraces in 1899 as of lacustrine origin,[3] but thought the lake had been drained by canyon-cutting below. The discovery of better-developed,

[1] Tarr, R. S. and Martin, Lawrence, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 161–166; Tarr, R. S., Professional Paper 64, U. S. Geol. Survey, 1909, pp. 125–137.

[2] McConnell, R. G., The Kluane Mining District, Geol. Survey of Canada, Ann. Rept., Vol. XVI, 1904, pp. 2A–4A.

[3] Brooks, A. H., Reconnaissance from Pyramid Harbor to Eagle City, Alaska, 21st Ann. Rept., U. S. Geol. Survey, Part II, 1901, p. 349.

higher beaches by McConnell in 1904, however, with mature forest hundreds of years old upon them, showing earlier lakes 300 feet deep, suggests that the glacier-dam explanation is correct and that there had also been still earlier advance and damming, perhaps contemporaneous with the earlier and greatest advance of the glaciers in Yakutat Bay, and elsewhere on the Pacific coast. Still earlier, but perhaps as a part of the same great advance, the glaciers had been so extended that the sites of these lakes were completely ice-covered. The north side of the St. Elias Range, therefore, serves to show the same glacial history as the south side:—(1) former great expansion of glaciers (perhaps with the 300 foot, forest-covered beaches associated with stages of ice-barrier lakes during retreat); (2) retreat of glaciers (since we cannot assume that the lakes stood at the 150 foot stages long enough for mature forest growth on the beaches above); (3) readvance a century or thereabouts ago with the formation of the 150 foot and lower beaches, since which there has been continuous retreat.

As shown by Wright[1] and by Reid,[2] Glacier Bay also furnishes evidence of a similar succession of recent events to that in Yakutat Bay and the north side of the St. Elias Range, and essentially contemporaneously with it. We may therefore assume that the last great expansion of the glaciers of the east end of the St. Elias Range was due to a cause which affected a mountain area with a northwest-southeast axis of at least 150 miles and a width of over 50 miles. But in Prince William Sound we find no evidence of recent glacier expansion. There was a period of great early expansion corresponding to that during which the Yakutat Bay glaciers reached their greatest extent, completely filling the inlet; but this period was centuries ago. Mature forest now extends up to the fronts of some of the glaciers of Prince William Sound and nearly up to the fronts of others. They are, therefore, now almost as greatly expanded as at any time within a century or two; and Columbia Glacier is greater than it has been for a long time, for both at its sides and front it is encroaching on the mature forest. It is possible that these glaciers are now in a stage of advance comparable to that of the Yakutat Bay glaciers a century or more ago.

Three possible explanations suggest themselves to account for this recent pronounced, but brief, advance of the Yakutat Bay glaciers and those in the two neighboring regions cited. The first of these is the diastrophic hypothesis, that uplift took place and the mountains were raised so much higher that the glaciers were thereby given greater snow supply. This hypothesis fails in one vital respect; by it we cannot account for the brevity of the advance and for the subsequent recession. It may, therefore, be dismissed as inadequate to account for the phenomena, as is the complementary hypothesis of submergence proposed by Reid to account for the retreat of the Glacier Bay ice tongues.

The second hypothesis is the vague one of climatic change, making no attempt to explain why it came, how it operated, or what its exact nature was, but merely for some unknown reason an increase in snowfall, or change in temperature, came to these mountains and that ultimately the glaciers flowing from them responded by advancing; then came diminution in snowfall, or increase in temperature, and accompanying recession. Lacking knowledge of the amount and nature of the snowfall and the temperature of the critical parts of the Alaskan coast, as we do, it is impossible to apply to this theory the

[1] Wright, G. F., Ice Age in North America, New York, 1891, pp. 51-57.

[2] Reid, H. F., Studies of Muir Glacier, Alaska, Nat. Geog. Mag., Vol. IV, 1892, pp. 34-40; Glacier Bay and Its Glaciers, 16th Ann. Rept., U. S. Geol. Survey, Part I, 1896, pp. 438-440.

test of actual climatic records. It would certainly seem remarkable, though, if after a long period during which the precipitation was not sufficient to push the glaciers out as far as their present termini, there came a sufficient increase in precipitation to push glacier fronts forward 15 or 20 miles or more, and then after a brief interval, sufficient diminution in precipitation, or increase of temperature, to allow the glacier fronts to recede 15 or 20 miles. Such a great climatic change would of itself be so remarkable that we ought to at least see a sufficient reason for it. Failing that, the hypothesis seems improbable, and the improbability is increased by the discovery of the fact that the change in the glacier fronts of Yakutat Bay, and those north of the St. Elias Range and in Glacier Bay, have not been paralleled a few hundred miles to the northwest, in Prince William Sound, which lies in the same general climatic belt and whose heavy precipitation is induced by the same meteorological conditions that cause the heavy precipitation in the Yakutat and Glacier Bay regions. While we do not feel that in this case, as in the case of the hypothesis of uplift, we can dismiss the hypothesis as inadequate, we nevertheless are of the opinion that it is highly improbable and to be resorted to only when no other more adequate explanation is possible.

The third hypothesis is that of earthquake shaking, which at first thought, might seem both extravagant and impossible, but in favor of which, nevertheless, much can be said. By this hypothesis it is assumed that a century or more ago there was a great and long-continued shaking of the snow-covered ranges of the mountain region from Yakutat Bay to Glacier Bay, perhaps in a series of years, with repeated great earthquake shocks. By this shaking and avalanche overloading of the glaciers a wave of advance was started, and in consequence of each great shock the advance was continued, until the glaciers reached their outermost positions; then, with cessation of vigorous shaking, the glaciers receded. The earthquake record unfortunately does not go back quite far enough for correlation as to this possibility, the earliest recorded earthquake which we have seen even mentioned having occurred in 1788.

This hypothesis has the one great merit that the other hypotheses lack: it would account for all the phenomena. By it we can explain the local nature of the advance, its absence to the northwest, and even the present advanced position of Columbia Glacier. The hypothesis would also account for the briefness of the advance, and for the subsequent rapid recession. There appears to be but one important objection to the earthquake hypothesis, and that is the stupendous response to it by the glaciers. It may be said, however, that the same objection applies also to all other hypotheses. There certainly was a great advance, difficult to explain by any of the hypotheses, and perhaps no more difficult under this one than under the others. In 1906 the senior author was inclined to believe this hypothesis so improbable that he resorted to "the vague hypothesis of climatic variations," but a further consideration of the problem, and especially the remarkable advance of Hidden Glacier, whose front was pushed 2 miles farther out in 1909 than it was in 1906, has led both of us to the belief that the earthquake hypothesis is not so utterly improbable as to warrant discarding it. It is certainly conceivable that a series of severe shakings, like those of 1899, repeated through successive years, might cause even so great an advance as the 15 to 20 mile advance of Hidden and Nunatak Glaciers. When we know the nature of the response that the Hubbard and Nunatak Glaciers are going to make to the earthquake shaking of 1899 we may be better able to consider this hypothesis. If they do not advance greatly we should consider it far less

probable than we do now; but if their fronts are pushed forward we should be inclined to consider this the most rational of the hypotheses proposed.

It is important to note that the former great advance of Hidden and Nunatak Glaciers 15 to 20 miles beyond their present termini is not so remarkable as may seem at first thought. If, for instance, Hubbard and Turner Glaciers should be pushed out two or three miles, which, in view of the recent advance of the much smaller and less active Hidden Glacier, is certainly a conceivable result of advance due to earthquake shaking, the entire Russell Fiord and Nunatak Fiord would be transformed to a lake by an ice dam at the head of Disenchantment Bay. The tideless, fresh water of this lake would cause less recession in Nunatak Glacier than the tidal salt water now does. Moreover, the lake water would be much colder than the waters of the fiord, which are influenced by direct connection with the ocean. This coldness would be increased by the icebergs from Nunatak Glacier, which could no longer escape to the Pacific, and by those from Hidden, Variegated, Orange, and perhaps smaller glaciers which would become tidal. It is even conceivable that the lake waters would be so filled with floating ice that they would remain frozen well into the summer.

The effect of all these conditions would be to aid in the extension of Nunatak and Hidden Glaciers in two very important ways: (1) by greatly checking the discharge of icebergs from the ice fronts, (2) by lowering the mean annual temperature of the region near by, and thus diminishing ablation and increasing snowfall. Thus a thrust which pushed Hubbard and Nunatak Glaciers forward a distance of only a few miles might result in a far greater extension of Hidden and Nunatak Glaciers. A part of this extension would need no other impelling cause than that which now exists; for with lower mean annual temperature, and lessened iceberg discharge the fronts of the glaciers would necessarily move forward beyond the position which they now hold as a balance between supply and wastage from ablation and iceberg discharge. When the ice dam across Russell Fiord was removed by recession of Hubbard Glacier, the return of the tidal water to the margin of the advanced Nunatak Glacier would inaugurate recession at a more rapid rate, and if the cause for the advance had by that time died out, the glaciers might be expected to recede rapidly.

These considerations are of course as applicable to an advance by other causes as to an advance in response to earthquake shaking. But since these associated phenomena make it clear that so great an advance as that of Nunatak and Hidden Glaciers does not necessarily mean an actual forward thrust of a glacier for 15 or 20 miles by the sole and unaided cause of response to earthquake shaking, they remove some of the objection to this hypothesis on the score of the enormous forward movement which actually occurred. Only a part of the advance is assigned by this hypothesis to the direct thrust; the remainder is ascribed to the result of accessory conditions indirectly the result of advance under the impulse of earthquake shaking. With this advance, and until further evidence is available, we believe that the earthquake hypothesis has basis for retention as a possible explanation of the remarkable advance of the Yakutat Bay glaciers, from which these glaciers were withdrawing when the earthquakes of 1899 gave rise to conditions by which new advance was necessitated in some of them, and perhaps, ultimately, in all the more important ones.

CHAPTER XI

GLACIATION OF THE YAKUTAT BAY REGION

Introductory

In the preceding chapters the main theme has been the consideration of the characteristics of individual glaciers, with only incidental reference to relationships and to more general problems, the single exception being the discussion, in Chapter X, of the problems presented by the advancing glaciers. There are, however, some features common to many glaciers, some phenomena and problems relating to the region as a whole, and some of general interest because of application to other regions. It is the purpose of this chapter to briefly treat these phenomena and problems topically, introducing such repetition of statements previously made as is necessary for clear individual presentation. The first group of phenomena and problems relate to the characteristics of the glaciers themselves; the second to special problems clearly illustrated by certain of these glaciers; the third to resemblances between the existing glacial phenomena of Yakutat Bay and regions of former glaciation; and the fourth to the glacial history of the region.

Characteristics of the Glaciers

As has been pointed out specifically in Chapter II, and as is clear from the descriptions in the succeeding chapters there are many different kinds of glaciers near Yakutat Bay, differing greatly in size, rate of flow, and a multitude of other characteristics. No matter how different they may be, however, they are all alike in one important respect, —in some portion of their course, often in the middle, and sometimes throughout, they are normal valley glaciers. But the valley glacier portions grade down-stream to termini which present great differences in characteristics; and up-stream there are also notable differences among the many glaciers of the region. These variations will be more specifically stated under four headings, the cornice glacier, the valley glacier, the through glacier, and glacier termini.

The Cornice Glacier. The cornice glacier, common in glaciated mountain areas, is characteristically developed in the Yakutat Bay region in two relationships. First, where it is nourished by excess of snowfall on rock ledges or cliffs, being, therefore, due to favorable topographic inequality combined with such a climatic factor as a north exposure. This kind of shelf, or cornice glacier has conditions almost exactly similar to the normal valley glacier, of which it forms an incomplete type, with surface plan of great irregularity and some times with a short valley glacier projecting from it. Many of the glaciers of the Colorado, Montana, and Canadian Rocky Mountains are of this sort. They are often much larger in Alaska, and exist in hundreds in the mountains of the Yakutat Bay region, being larger on north than on south slopes, and in the irregular heads of valleys than in the smoothly-sculptured lower valley troughs,

where intense glacial erosion has removed the ledges. The second condition is where these cornice glaciers seem to be remnants of the thinner, shrinking valley glaciers of a region, and are not nourished where they rest but are slowly wasting away or remain nearly in balance between snowfall and wastage. Both types contain true glacial ice, whose blue or green may be seen in crevasses. They often move forward to cliff edges, where fragments tumble to valleys below. They grade into the snowbank without ice, on the one hand, and into the true valley glacier on the other.

The Valley Glacier. The Alaskan valley glacier is in no fundamental respect different from its counterpart in the Alps, Caucasus, Himalayas, and other mountain regions, and may, therefore, be considered normal. There are, of course, many differences in reservoir conditions, in number and nature of tributaries, in grade, in rate of flow, in extent of crevassing, in size, and in other less important respects. A few in the Yakutat Bay region have their sources in normal cirque reservoirs among lofty mountains; more flow from the steeply-inclined snowfield areas on precipitous mountain tops and slopes; some head on the broad flat divides of through glacier systems. There are so many glaciers that do not head in cirques, the snow supply is so heavy, and avalanches from valley heads and walls are so frequent that the bergschrund is not always well defined near Yakutat Bay.

The glaciers are usually enclosed between steeply-rising ice-eroded mountain slopes, from which both snow and rock are frequently avalanched upon their surfaces. The snow line lies from 2000 to 3000 feet above sea level, and in all the larger glaciers a part of the actively-moving ice stream lies above snow line, so that the névé region is likely to be concealed beneath the lower end of the snowfield. The glacier is supplied from upper reservoirs, from snowfall upon the glacier ice below these, from the avalanching of snow from steeply-rising enclosing mountain walls, both in and below the reservoir zone, and from tributaries. In this region of heavy snowfall the glaciers are so well supplied that even small tongues push their termini practically to sea level, while larger ones either enter the sea or spread out to the mountain base.

The valley glacier has the normal morainic load common to such glaciers. There are medial moraines here and there, and lateral moraines are almost universally present, while the lower ice layers are heavily débris-charged. In some of the glaciers conditions of rock supply and ablation are so favorable that a sheet of débris coats the lower glacier from side to side, forming the broad fields of ablation moraine that are more fully described and discussed in a later section. In the dissipator, lateral moraines become etched into relief and a marginal valley commonly develops on either margin of the glacier, in which streams flow and deposits of complex nature are being accumulated. These marginal conditions are absent in only a few cases, the three most pronounced being (1) where recent spasmodic advance has closed up marginal valleys previously existing, (2) where the glacier terminates in a cascade on the steep face of the valley wall, and (3) where the combination of ablation and iceberg discharge in an actively-moving valley-enclosed glacier, cuts back into the dissipator too rapidly for the development of marginal valleys, as is illustrated in Nunatak Glacier.

The deposits which would be left in these mountain valleys if the valley glaciers should melt out of them, would, about the termini, be closely like those of other valley glacier regions. There would be a fairly thin sheet of ground moraine, more or less well-defined medial and lateral moraine bands, and marginal valley deposits of varying character, extent,

and form, according to the conditions surrounding their accumulation and the length of time occupied by it. One notable difference from normal valley glacier deposits would be found in some of the valleys, in which a sheet of somewhat weathered, angular fragments would overspread the ground moraine. This, representing the ablation moraine, would form a thin sheet, varying in thickness and with a somewhat hummocky topography.

The Through Glacier. The through glacier condition has already been described with such fulness that little remains to be added here. It is a common condition in the Alaskan coast ranges to find mountain valleys so flooded with ice to the divides that one can travel by fairly even grade from the dissipator of a glacier up across a broad, flat divide-reservoir, then down to a dissipator in another valley, without being able to determine where the boundary line between the two glaciers is to be drawn. Such is the through glacier, occupying a through valley with low divide. In some cases more than two ice tongues extend from a through glacier divide area. Because of the extensive development of the through glacier condition there is a network of ice-flooded valleys, the full extent of which can only be inferred in the present state of our knowledge of the Alaskan coast ranges back from the coast.

The through glacier condition is certainly not confined to the Alaskan mountains, but it is more strikingly developed there than in any other region of which we have seen description. Its extensive development here is dependent upon the combination of three very favorable conditions, as follows:—(1) the fact that at a former period the glacier systems of the Alaskan coast were far more extensive than now; (2) the fact that this period of glacier expansion lasted long enough for extensive glacial erosion to lower many divides; and (3) the fact that, even in the present shrunken condition of the glaciers the snowfall is so heavy that both glacially-deepened valleys and lowered divides are still ice flooded. There is reason to believe that in the former period of expanded glaciers, and before the present valleys were so deepened by glacial erosion, ice currents streamed across many divides, and in different directions from now. Some of these divides were completely removed by erosion, but many were only partly removed; and in the latter case, as the volume of the ancient glaciers shrank, the time came when some of these lowered divide areas became the flat ice-divides of through glaciers. Doubtless in other cases, back among the mountains where, as Russell vividly describes it, the ice-flooded valley condition is so well developed, both valleys and divides are still so deeply buried beneath snow and ice that glaciers are streaming across divide areas and still lowering divides between mountain valleys.

One of the needs of Alaskan glacier study is the exploration and mapping of some of the through glacier systems, for at present we know very little about the conditions a few miles back from their termini. It will be interesting to know more about the source of the ice supply, though in all probability this varies greatly in different glaciers. Lying, as they do, in broad valleys among extensive, lofty, snow-covered mountains, it is certain that they receive important avalanche contributions from the enclosing mountains, as well as contributions from tributaries descending to them from the upper mountain valleys. In the larger, more active, through glaciers, these must furnish the main ice supply; but in some of the smaller, less active glaciers, like the Orange Glacier, and perhaps even the south arm of the Nunatak Glacier, a large part of the snow supply, and perhaps even the dominant part of it, apparently comes from the snow that falls on the broad divide and upon the glacier surface itself.

It is noteworthy that, while smaller glaciers are often covered with broad sheets of ablation moraine, through glaciers are not so characterized. In the active tidal glaciers, such as the Hubbard and Nunatak, this might be explained by the failure of ablation in connection with iceberg discharge; but in such glaciers as the Hidden and Fourth, where ablation lowers the terminus to a thin edge, and where the terminus lies in the same relation to sea level as smaller, neighboring moraine-covered glaciers, this explanation fails. The absence of moraine sheets in the lower glacier in such cases must be explained by failure of débris to be incorporated in the ice so extensively in the through glaciers as in the smaller valley glaciers. This lack of incorporated débris may be due to a combination of several causes, varying in relative importance in the different glaciers, among which are the following:—(a) the fact that a considerable part of the ice is supplied by snow falling directly on the glacier; (b) the breadth of the through glaciers, making it impossible for avalanches to spread completely across the glaciers as they may in narrower glaciers; (c) the relative weakness of tributaries by which the débris that they contribute is kept on one side of the main glacier by the greater force of its current; (d) the deep snows which mantle so large a proportion of the upper mountain slopes with so deep a cover that the mountain rocks are protected far more than the steep, bare slopes which border a large portion of the smaller valley glaciers.

Even where through glaciers emerge beyond the mountain valleys and expand in piedmont areas, as in the Malaspina Glacier, some of whose tributaries are probably through glaciers, the extent of the sheet of ablation moraine is proportionately less than that of the smaller glaciers. Thus, although there are such broad sheets of ablation moraine on the Malaspina as to have attracted wide attention, the most extensive of these are really marginal, and related to the lateral accumulations by the tributaries which supply the Malaspina ice. Extensive as these sheets of ablation moraine are on the Malaspina Glacier, when compared to the glacier system as a whole they are far less notable than the ablation moraine of the Atrevida-Lucia piedmont glacier system. The Malaspina and its tributaries are dominantly clear ice, with only relatively-small, peripheral, ablation moraine areas and larger lateral sheets; but the smaller Atrevida-Lucia piedmont glacier is completely moraine-covered, and the morainic sheet even spreads well up the valley portion of each tributary. Extensive morainic sheets are, therefore, not characteristic of the through glaciers, but rather of the smaller glaciers, being present even where they terminate within their mountain valleys, but being far more developed where their termini expand in piedmont areas outside of their mountain valleys.

The Glacier Termini. It is the lower portions of the Alaskan glaciers that have so far been most carefully studied, and it is these portions, also, that present the most varied and interesting phenomena, and those which supply most facts of value for comparison with glaciers of other regions, both existing and extinct. Among the Alaskan glaciers there are many different conditions in the termini, and as most, if not all of these are illustrated in the Yakutat Bay region, it seems worth while to briefly state in the following paragraphs some of the more notable differences observed.

The termini of many glaciers, probably the majority, are at their junction with larger glaciers to which they are tributary. The conditions here vary greatly according to the relative size of tributary and main glacier and to the position and direction at which they enter. In the vast majority of cases these tributaries cease at once to have notice-

able influence, the ice and débris supply which they contribute being so small a proportion of the whole that they are completely dominated. Occasionally a tributary is so nearly equal to the main stream that it maintains individuality, both in its ice currents and in its morainic areas, clear to the glacier terminus; and now and then, where two large glaciers unite, one is so much the stronger that, as in Hubbard, Nunatak, and Fourth Glaciers, both the ice current and the moraine of one are pushed far over to one side by the greater force of the other. On the other extreme we find many cases of glaciers, especially those cascading from small hanging valleys, which barely coalesce with the main glacier, and sometimes just fall short of doing so. In such cases local enlargement of the lateral moraine of the main glacier occurs at and just below the point of incoming of the tributary, but otherwise no noticeable influence is caused by it.

Some glaciers terminate in hanging valleys, at varying distances back from the lips. Such glaciers are often moraine-covered at their ends, and, if they halt long enough, they build terminal moraines. From the terminus one or more streams issue, heavily charged with débris, which in some cases, where the valley grade is steep enough, is borne on beyond the lip of the hanging valley, but in other cases is partly deposited on the floor of the hanging valley, forming a more or less perfect outwash gravel plain. In all cases a large porportion of the débris is borne on to the lip of the valley, thence down the steepened slope to the flatter slope at its base, or to the sea, where deposit is extensive. Here are built steep alluvial fans on valley sides, and large, steep deltas where the steepened slope descends to the fiord. The great volume of water, with its sediment load and its steep grade has invariably cut a gorge below the hanging valley lip which it is rapidly deepening; but the time since these streams have been flowing is so brief that the gorges are rarely deep or broad. They increase in depth and breadth toward the outer portions of the fiord, as one would expect in view of the fact that the steepened slopes there have longer been exposed to the erosive action of the stream; but this increase is less striking than might be expected, since the glaciers in these hanging valleys are smaller than those in the inner portions of the fiord and the water supplied by their melting is therefore less in quantity.

A few hanging valley glaciers extend to the very lip of the valley, and some extend even beyond it, terminating as cascading glaciers. In these cases there is a smaller alluvial fan or delta at the base of the steepened slope, and a less developed gorge has been cut in the steepened slope. Indeed, the cascading glaciers may have no gorge whatsoever, their streams flowing along several courses on the surface of the steepened slope. These glaciers usually have a more or less well-defined moraine at the cliff base in which is incorporated snow and ice blocks that have fallen from the steeply perched ice terminus. Being only recently disconnected from the main glacier, and having usually only a brief duration in steeply-perched positions, the cascading glaciers do not commonly form extensive deposits or leave striking records of their former presence; and such records as they do leave are soon buried beneath the growing alluvial fans.

Fairly good sized glaciers that terminate in valleys near sea level, where ablation is rapid, have made much more extensive deposits than in any of the preceding cases. If the glacier fronts were stationary, terminal moraines would doubtless develop; but, since the glacier history of Yakutat Bay has been one of general recession, such deposits are not common, though they have been found in a few places, such as Calahonda valley. The receding glacier contributes most of its débris to streams, leaving only thin deposits

of coarser fragments where they were brought by the glacier, and especially in marginal deposits. The glacial streams issue with great volume in this climate, and they bear vast quantities of sediment, varying from clay to coarse gravel and small bowlders. Where the grades of the valleys are steep enough, much of this débris is borne to the fiord, but, in those valleys with flatter grade, outwash gravels accumulate; though even in these cases extensive deltas are built in the fiord, and vast quantities of clay are contributed to the deposits being laid down on the fiord bottom. At times, as in the case of Hidden Glacier, outwash gravel plain and delta coalesce, and from the glacier front to the sea there is a continuous outwash gravel plain, growing finer in texture toward the sea, and beyond mid-tide level consisting mainly of clay. Such deposits, extending from one side of the valley to the other, overspread and bury from view those deposits that were previously made beneath the glacier, or by its melting; and, in fact, they sometimes bury the glacier end itself, giving rise to extensive buried ice blocks.

Some of the smaller glaciers that terminate in the position here being considered have their ends covered by ablation moraine and are in a stagnant state; and in some cases the lower ends are even disconnected from the upper part of the glaciers. Under these conditions the rate of ablation is decreased, the extent of outwash gravel deposit is diminished, and neither glacier ends nor glacier deposits are extensively buried beneath the gravels. In such cases the record of former presence of glaciers will include a sheet of slightly-irregular moraine, including a large percentage of angular, frost-riven blocks. One might expect to find eskers in such situations, but not where the recession of glaciers is so closely followed by gravel deposit as to cover both previous deposits and glacier ends. In the latter case, however, conditions are favorable for extensive development of kame topography when the buried ice blocks melt and the overlying gravels slump irregularly.

The Nunatak Glacier is the only case in this region of a tidal glacier which terminates in a mountain valley which it completely occupies from side to side. Ablation is rapid and iceberg discharge active, the two causes together sufficing to cause rapid recession of the glacier front. Marginal drainage is not developed, but subglacial or englacial streams pour volumes of water into the fiord in front of the glacier. Doubtless these are giving rise to extensive sedimentation in the neighboring fiord, but owing to the rapid recession of the glacier front these deposits are spread over a wide area. Should the front halt for a time the vast quantities of sediment that are doubtless pouring out from the ice front would build notable terminal deposits. In the deposits that are accumulating both fine and coarse material are of necessity included, thus differing notably from the outwash plain deposits in which the clay element is almost completely absent. These submarine deposits must also differ from those of the land streams in the presence of large rock fragments dropped from the glacier front or from the icebergs; but the number of these fragments is not great, for the larger proportion of the material that the ice bears to its front is carried away in the icebergs. This deposit should be assorted throughout, for in addition to the fact of its deposit in water, in which there are strong tidal currents, it is brought to its place of accumulation by rapid streams, whose continued progress is arrested by the mass of standing fiord water.

The other two tidal glaciers—Turner and Hubbard—differ from the Nunatak in one very notable respect. They extend beyond enclosing mountain walls and spread near their terminus. Were the fiord water absent, these glaciers would doubtless develop

expanded piedmont bulbs; but the attack of the sea water and the rapid dispersal of ice that falls into the fiord from the glacier front, succeeds in truncating the bulbs, so that they are only partially developed. In those portions of the ice front from which icebergs are discharged the conditions are similar to those of Nunatak Glacier, and the remarks made about the deposits there apply in all essential particulars here also. Along the margins, however, the conditions are different, for in their lateral spreading both the Turner and the Hubbard Glaciers have extended so far as to give rise to small areas of stagnant or partially stagnant ice on either margin. Along these less active areas marginal streams flow and each has built a small alluvial fan.

A number of glaciers, including several of small size, expand at the mouths of their valleys giving rise to piedmont bulbs, or bulb glaciers. These are sometimes in larger valleys, like Variegated Glacier in the Russell Fiord trough, but more commonly are at the mountain front, like Galiano Glacier. Where two or more glaciers descend from neighboring valleys, and expand sufficiently, the piedmont bulbs may coalesce and form a piedmont glacier, as is illustrated by the Atrevida-Lucia, and, far better, by the Malaspina Glacier, formed by the union of several large glacier bulbs, and the type case of the piedmont glacier, made known to us by Russell's researches. The piedmont glaciers and the piedmont bulbs have many peculiar and interesting features, including marginal deposits of varied character, ablation moraines, buried ice blocks, and outwash gravel deposits, fully discussed in later pages.

Special Problems

Among the glacial phenomena some warrant more detailed discussion. These relate mainly to the piedmont condition, to some of the phenomena associated with other glaciers, and to the work of the formerly more extensive ice tongues.

The Piedmont Bulbs. The phenomenon of spreading of glaciers on emergence from valleys is common in the Alaskan region. Indeed, it is universal wherever the glaciers extend beyond the valleys which confine their upper portions, whether the escape be into a broader valley, as in the Variegated Glacier, or beyond the mountain front, as in the Atrevida. This spreading to the piedmont bulb condition is evidently due to lateral flowage where an ice supply is released from lateral confinement, and it bears a very close resemblance to the flowage of wax or other viscous substance. We interpret it as a phenomenon of viscous flowage of unconfined ice.

In all the cases which we have studied, the phenomenon of the piedmont bulb expansion is the product of a previous stage. We cannot, therefore, give a description of the piedmont condition at the time of formation. We infer, however, that the ice of the entire piedmont area was then in motion, though with great diminution in velocity, toward the periphery, reaching stagnation and perhaps being almost motionless around the margin, and especially in the lateral margins. Nor were we able to describe the ice currents in such a bulb. Judging from the great morainic swirls on the Malaspina Glacier, and the crescentic banded moraines on the Variegated Glacier bulb, the ice currents are complex and peculiar and a determination of their nature would be a matter of great interest. The currents seem to give rise to a broadening and spreading of medial and lateral moraine bands; but our knowledge of their nature and behavior is so slight that we are unable at present to go farther than we have done in our description and interpretation of the morainic phenomena of the Variegated Glacier.

One might expect important influence from uprising bottom currents, and, as is shown in a succeeding section, there is evidence that there is such influence in the production of interior flats. But, in general, there does not seem to be much effect from uprising of ice, for the moraines which cover the piedmont bulbs are prevailingly of angular material, such as falls upon glacier surfaces and in glacier reservoirs, not such as has undergone the scouring necessary in bottom ice layers. It is hoped that further study of Alaskan glaciers may yield facts which we lack at present concerning the nature of ice motion in piedmont bulb areas.

The piedmont bulb areas which we have so far studied have at some previous stage of greater expansion spread to their present extent, and doubtless farther. Whether this expansion was one of rapid or slow advance we do not know, nor can we assign to it an exact date. In the cases of some, however, notably the Lucia, Atrevida and Galiano Glaciers in Yakutat Bay, we know that the period of stagnation following the advance began not less than half a century ago, for trees of that age have grown on the stagnant outer portions of these bulbs. A similar statement applies to the piedmont bulbs of Allen, Heney and Miles Glaciers in the lower Copper River valley, though they may possibly be a little younger. Malaspina Glacier maintains some activity throughout most of its area, being motionless only in a few parts where the spreading has been greatest.

Following the expansion of the piedmont bulbs came a cessation of supply sufficient to maintain the bulb portion, and normally one might expect rapid destruction by ablation. In these cases, however, such destruction has been prevented by the concentration of moraine on the surface of the bulbs so protecting the ice as to greatly reduce the rate of ablation. In fact, in some of the outer portions of the bulbs the moraine has become so thick that ablation has almost ceased. From the forest-covered bulbs of the Lucia and Atrevida Glaciers, for instance, only trickling streams of water ordinarily emerge, and the stability of the buried ice is so great that dense, continuous forest growth covers the ice. Also, where alluvial deposits have been laid down on the ice, as in the bulb of the Galiano Glacier, the rate of melting of the buried ice must be exceedingly slow. In such cases it is probable that scores of years, and perhaps even a century or two, will be required to remove the ice from beneath its protective covering of moraine and alluvial deposit. The exceedingly slow rate at which the moraine-covered ice melts, and the fact that Malaspina Glacier inside the moraine-covered portion is not greatly lowered, is interpreted as evidence that this glacier is still in motion throughout most of its area; it promises a rich field for detailed study and one that will probably throw much light on the movement of ice in piedmont glacier bulbs.

Ablation Moraines. In their present wasted stage the piedmont bulbs present several interesting phenomena due to the progress of wastage. One of them is the presence of extensive sheets of moraine, some of which extend up the valley glacier portion. Similar sheets occur on small glaciers which are entirely confined to mountain valleys. Although differing greatly in detail of form, extent, and composition from place to place, the ablation moraines throughout the region have the same general characteristics. Some of the details peculiar to individual glaciers have been considered in the chapters discussing these glaciers; it now remains to consider the ablation moraine in general.

On any individual glacier the moraine sheet is found to vary greatly in thickness from place to place. The broadest areas of thickest moraine are the peripheral portions where stagnation has lasted longest and ablation has, therefore, been most effective. Here

the moraine may be from 5 to 10 feet deep. Marginal bands of deeper moraine also extend up the lateral margins of the valley glaciers, marking the site of the lateral moraine accumulation. There are crescentic bands of thicker moraine in the piedmont bulbs, giving rise to ridges because of retarded ablation, while crescentic valleys lie between the ridges where the moraine is thinner. As ablation proceeds the position of the ridges and valleys change, for when the ridges become too steep the moraine slides into the depressions. Speaking generally, there is a gradational decrease in thickness of ablation moraine from the periphery of the piedmont bulb to the part of the valley glacier where the ablation moraine disappears, a point which varies greatly from glacier to glacier, but which is always in the area of the dissipator, and well below the snow line. This decrease in thickness is locally interrupted by areas of thickening on ridge tops and of thinning in valleys and on steep ridge slopes.

There are variations in nature of material composing the ablation moraine. Speaking generally it is made up of frost-riven angular fragments of rock of the kind enclosing the upper glacier, and often includes bowlders of huge size. Scratched stones are also found, though not commonly, and there are even areas of clay and waterwashed gravel, for short streams flow on the moraine, and pools are not uncommon. There are bands, often crescentic, in which rocks of one kind so predominate as to give rise to bands of color on the glacier surface, as in the Variegated Glacier. With our lack of knowledge of the behavior of ice currents in spreading glaciers, and our ignorance of the bed rock in most of the enclosing valley walls, it is not possible to offer a definite explanation of these variations. They are without question due to the flow of the ice in the piedmont bulb, distributing the load which the ice bears and which is brought into prominence by ablation; but the nature of the process is not yet clear to us.

On the thicker, outer portion of the ablation moraine the soil has such stability that vegetation grows luxuriantly and one often needs to study closely to determine where the land forest ends and the glacier forest begins. It seems that from 5 to 10 feet of moraine on ice in this climate is sufficient to give rise to such a condition of stability as to permit practically uninterrupted forest growth with spruce, hemlock and cottonwood trees like those on land. The density and maturity of the forest growth progressively diminishes from the peripheral zone, and near its inner margin there are abundant signs of the struggle to which plants are subjected when growing in a slumping soil. The average age of the plants, here, mainly, if not entirely, alders and willows, may be from ten to twenty years, and the majority of individuals are healthy and undisturbed. But areas occur where for some reason ablation is locally more active, and there plants are found overturned, others with their roots partly uncovered, others partly inclined; in fact all stages in plant destruction are to be seen, and perhaps on the slopes of some basin containing a pool, or where a moulin has opened in the moraine, the ice itself may even be seen. Looking down upon a forest-covered ablation moraine from some favorable viewpoint, numerous areas of this sort may be seen. They seem to be related to the development of ice drainage in the vegetation-covered parts of the piedmont bulb; but they occupy only a small proportion of the vegetation-covered part of the moraine and are mainly confined to the inner part of the zone. Even here, however, the condition of soil stability is in general sufficient to permit the growth of mature alders.

From the zone of occasional slumping there is a rapid gradation to the barren zone of the ablation moraine, which is a true desert, almost devoid of life. In this gradational

belt, vegetation is struggling against increasing odds until, finally, plant growth becomes impossible because of the excessive instability of the soil. The seeds of the alder are taking root all over the ablation moraine within the zone of vegetation; and, near the inner border of that zone, occasional plants may succeed in maintaining growth for a year, or two, or three in especially stable areas; but, ere long, even the most stable soil yields to the undermining and the shrubs perish. One finds plants with green leaves, but with nearly all the roots exposed; others dead; and some, overturned and buried, with their roots in the air and their branches under ground, sending out roots from the stems. It is a hopeless struggle, and at a distance of a few hundred yards from the inner alder zone few, if any, plants are found. Grasses and annual plants, and even lichens on the larger bowlders no longer find it possible to survive the constant rolling and sliding to which the morainic soil is subjected.

When subjected to advance, as Atrevida, Galiano and Marvine Glaciers were, even the zone of stability temporarily becomes impossible for plant growth. Many trees and bushes are destroyed by the direct thrust and breaking of the ice; but many more are destroyed by the sliding of the soil into the newly-formed crevasses. By these two means great windrows of dead alder and other trees were caused by the 1906 advance.

On the ablation moraine desert, where no plant life can find a foothold, the evidence of instability due to rapid ablation is everywhere present. Here the average depth of moraine cannot be more than two or three feet, and ice is to be seen here and there. The surface is exceedingly hummocky and in whatever direction one travels it is necessary to go up hill and down, rising 50 to 150 feet from valley bottom to hummock crest. There seems no system in detail, though the topography is, in general, a series of roughly circular or elliptical kettles with enclosing ridges of varying heights. From the steeper slopes, bowlders are constantly sliding and when one takes a step on such a slope he may start an avalanche of stones, or he may himself slide down the descent when his foot comes in contact with the thinly-veneered ice slope. Even on the more level parts, bowlders are often perched in such unstable positions on hidden ice pedestals that a mere touch may overturn them. Everywhere instability is evident and there is complete demonstration of the rapid progress of ablation; but when viewed from a distance the ablation moraine seems only a barren waste of débris with little or no ice to be seen.

In spite of the rapid ablation on the moraine-covered ice, there is little flowing water to be seen, excepting where it emerges in great torrents from the glacier margins. There is a great abundance of trickling rills down the steeper slopes, and probably also beneath the moraine on the lesser slopes, but this drainage is usually toward small enclosed basins from which escape is found into the ice. Now and then one of the kettles has no outlet and then a pool or pond is formed, but these are relatively uncommon. The nature of the outlets of the kettles is generally hidden from view by the débris which has slid into them from the margins; but not uncommonly one can hear the falling water as it cascades into hidden moulins. Only in the larger ice valleys are there large moulins. There the ice drainage follows small rills into many small kettles, thence into the glacier and probably soon to its bed. Evidence of the presence of at least some englacial streams was seen in several places where tunnel ends were exposed to view by the melting of the glaciers. Since the sliding of débris from the ridges into the kettles transforms areas of thicker moraine to areas of thinner deposit, and since the largest fragments, at least,

cannot be carried out of the kettles into which they slide, we assume that kettle areas of one period become hummock crests in others, and that one-time ridges later develop into kettles. If this is so, the position of the glacial drainage is constantly changing in detail.

By the ablation and the consequent drainage some of the moraine is carried off by the streams, but the proportion thus removed cannot be great, otherwise it would be impossible for such extensive sheets of ablation moraine to accumulate. Most of the rills are fairly clear and free from sediment, and since the streams do not ordinarily attain large size or flow for more than a few score yards before descending into the glacier, the load which they bear cannot be great. They certainly could not carry bowlders and large pebbles, and the evidence is convincing that they do not carry large quantities of even the finer material. The glacial torrents which emerge from the margins of the piedmont bulbs are, however, heavily burdened with sediment; and, although some is doubtless contributed from the ablation moraine, we infer that much the greater proportion is derived from the lower layers of the glacier.

The extensive sheets of ablation moraine on the piedmont bulbs have two important effects. In the first place they prevent most of the local influence on climate which such extensive ice sheets would normally exert; and consequently dense vegetation can grow up to the glacier margin, and even on its surface where soil stability is sufficient. Secondly, they retard the melting of the glaciers, especially the outer parts, so that such ice sheets persist long after they would if not moraine covered. It would be interesting to know the rate of lowering of such covered glaciers, and it is our hope and expectation to undertake some such measurement, or at least to provide a beginning for measurement. Until, however, we have some carefully-run lines of levels, and some knowledge of ground temperature on and near the buried ice masses there can be little gained from a discussion of the probable duration of the buried stagnant ice. That it melts very slowly and lasts a long time is abundantly proved.

Moraine-covered glaciers are not confined to Alaska, though it is from the Alaskan glaciers that we have obtained our fullest knowledge of this condition. They are present in the Himalayas, and there are ice tongues resembling them, though without piedmont bulbs, in the Alps. In both these cases, and, in fact, in general, and even in Alaska, the clear ice glacier is the normal, and the moraine-covered glacier the exception. Our knowledge of the distribution of this type of glacier and of the surrounding conditions is not sufficient for anything like a final consideration of its cause. Yet there are some facts already observed in the Alaskan region which contribute toward an explanation of the phenomenon. Among these one of the most notable is the fact that the larger glaciers are less liable to the condition than the smaller; and another noteworthy fact is that small, valley-enclosed glaciers are commonly moraine-covered whereas the large through glaciers are not. A third noteworthy fact is that the ablation moraine is prevailingly made of angular fragments such as are contributed by avalanches. The inference that we draw from these three facts is that in certain glaciers whose walls are steep enough, whose width is small enough, and whose enclosure is sufficiently complete, enough débris is avalanched into the reservoir and out upon the valley tongues to provide material with which ultimately, through ablation, to clothe the entire glacier surface with a sheet of ablation moraine. In broad glaciers and in through glaciers all the necessary conditions are not present and consequently complete covering by ablation moraine becomes impossible.

Doubtless there are other factors which must be taken into account in a complete analysis of the phenomenon, and some of these are already evident. For instance, there must surely be steep slopes; but those are common enough and are present even in the valleys of glaciers which have no ablation moraine. Sufficiently steep slopes are almost invariably insured where dwindling glaciers occupy the beds of valleys overdeepened during a previous stage of greater expansion. Without steep slopes, such as glacial erosion provides, extensive ablation moraines are not possible; but that their presence alone is not a sufficient cause is abundantly proved by the absence of such moraines in valleys whose walls are as steep as those of moraine-covered glaciers. Another factor that is certainly important is friability of rock. The best ablation moraines of the Yakutat Bay region are in the friable Yakutat shales, sandstones, and conglomerates; but they are also present on glaciers, like the Variegated, which descend through valleys in crystalline rocks. It is probable, however, that extensive ablation moraines, such as those of Yakutat Bay, could not develop on glaciers whose entire course was through valleys enclosed in massive gneiss or granite. We are inclined to believe that the absence of ablation moraine on many small glaciers, enclosed between steeply-rising valley walls, like some of the Alpine glaciers, is due mainly to the stability of the valley wall rock.

A third possible factor, naturally suggested in this region, is that of earthquake avalanching. Valley walls are greatly steepened by glacial erosion, especially in the weaker friable rocks, and by shrinking of the glaciers these too-steep walls are exposed to subsequent weathering. One finds abundant evidence on every hand that weathering is working rapidly to reduce such slopes, for avalanching is commonly observed, and the rock is crossed by rifts and gashes, where weathering is preparing masses to slide down the steepened slopes. Sometimes these gashes in the upper part of the steepened valley slope are so wide that one cannot cross them, and scores of yards in length. When an earthquake comes there are many such masses ready to fall under the impulse of the associated shaking; and we have abundant evidence that great numbers of such masses did fall during the earthquakes of 1899. Such masses may often be of sufficient size to spread completely across a valley glacier. If a glacier bearing such a load is caused to advance by earthquake avalanching, we have a means of quickly moving down the valley the rock masses which the avalanching supplied to the glaciers; and if the piedmont bulbs themselves are the product of a great advance under the earthquake impulse, as is possible, the broad sheets of ablation moraine which cover them may in part be due to this cause. Earthquake shaking would have the double effect of quickly rolling to the glacier surface a larger amount of material than mere weathering would supply in the same time, and of contributing it in such great individual avalanches that it could spread farther out on the glaciers than would be common in those avalanches caused by weathering alone. The aid of earthquakes is not essential, but it simplifies the process and is perhaps an important factor. It would also explain the fact which we observed in Atrevida Glacier, that the area of ablation moraine desert extended farther up the glacier in 1909 than it did in 1905.

Interior Flats. In the description of Variegated Glacier it is shown that there is a level area, crescentic in form, and bordered by moraine-covered ice which rises steeply 100 to 150 feet above the plain. This level area is evidently a part of the glacier which for some reason had a thinner moraine cover and, therefore, wasted so much faster than the neighboring parts of the ice that it had, by 1905, become the seat of alluvial

deposit from glacial streams emerging from the Variegated. In 1905 ice still existed beneath the flat, and alluvial deposit upon it was rapid. The advance of the Variegated Glacier in 1906 overrode a small part of the flat and destroyed one of the streams that had flowed out upon it in 1905. In 1909 the condition of the flat was not greatly different from the 1906 condition.

The interior flat seems to be a fairly constant feature of piedmont bulb glaciers, being developed just outside the mountain front. In the Butler Glacier, for example, the interior flat is just within a crescentic area of moraine-covered ice now completely detached from the glacier, and representing the last stages in the destruction of a piedmont bulb. Here no ice seems to exist beneath the former interior flat.

Galiano Glacier also has an interior flat area. The valley glacier expands beyond the mountain front and appears to terminate in a moraine-covered ice cliff; but beyond this is an area of lowland, which in 1890–91 was the seat of extensive, alluvial-fan deposit. The advance of Galiano Glacier prior to 1905 destroyed this flat by disturbing the ice beneath it, and also extended beyond the flat, raising the moraine-covered ice in a series of hummocks over a broad area. Evidently Galiano Glacier has much the same condition as Variegated Glacier except that the buried ice is thicker, and the alluvial fan on the interior flat was larger.

A perfect interior flat exists in the piedmont bulb of Allen Glacier in the Copper River valley, but here it is less advanced than in the cases already described, for alluvial deposit upon it has only just begun, and the ablation moraine on the inner side is less notably developed. It is possible that the lake in front of the Miles Glacier is developed on the site of an interior flat in that glacier. The Heney Glacier on Copper River also has an interior flat.

There is no well developed interior flat on the Lucia Glacier, though in 1905 and 1906 there was a small area of clear ice with a lower surface, in the midst of the ablation moraine, and just where the glacier began to expand notably. It was half or three quarters to a mile long and a quarter of a mile broad in its widest part, and offered the easiest route over the otherwise moraine-covered, hummocky surface of the glacier between Terrace Point and Floral Pass. This was being destroyed by the 1909 advance.

In 1905 and 1906 there was no interior flat area on Atrevida Glacier, but a very perfect, well-defined area of clear ice was introduced by the 1906 advance and, by the progress of ablation in the interval between its formation and our next visit in 1909, its clear surface was lowered well below the level of the surrounding area of moraine-covered ice. As a result of the 1906 advance a similar area, though not as well defined because not so thoroughly moraine enclosed, developed on Variegated Glacier some distance inside the previously-existing flat.

The interior flat is so frequent an associate of the piedmont bulb condition that we believe it to be due to the operation of some general cause. The nature of this cause is suggested by the development of the area of clear ice in Atrevida Glacier during the 1906 advance. As stated in the discussion of this phenomenon there is strong reason for considering it the result of the upflow of clear ice from below, when, during advance, the flowage of the ice is retarded by the resistance of the stagnant, partly rigid, outer portion of the piedmont bulb.

For the phenomenon we propose the following working hypothesis, whose fuller discussion may be postponed until we have a larger body of observational data from a

greater series of glaciers. When a piedmont bulb is formed, the ice spreads as long as the supply is maintained, and, as ablation proceeds, the surface of the bulb becomes coated with moraine, if the supply of incorporated débris is sufficient. Under such conditions an interior flat should not be expected. But if a period of stagnation is succeeded by readvance, and especially if the advance be rapid, the spreading of the ice is interfered with by the stagnant outer part of the piedmont bulb, and, at the appropriate place, rising of lower ice occurs, introducing an area of clear ice where none existed previously. According to the thickness of the outer glacier, its degree of stagnation, and the rate and duration of the advance, the position and extent of the interior flat will vary. Thus with two advances of different intensity, or with different ice conditions, two interior flat areas, in different portions, may develop in the same glacier, as has occurred in the Variegated Glacier.

Marginal Deposits. The deposits accumulating around the margins of the Alaskan piedmont glaciers are complex and interesting. The most notable fact is that they are, in the main, water-laid. Doubtless by the melting of the ice a veneer of ground moraine is being accumulated on the site of the glaciers; doubtless, also, upon this is accumulated a veneer of angular débris from the ablation moraine; and doubtless there are subglacial deposits of water-laid material, such as eskers; but in the areas occupied by the piedmont bulbs these deposits are for the most part masked by deposits from the water which issues from the wasting glaciers. The burial of these deposits is greatly aided by the slowness of wasting of the moraine-covered glaciers, as a result of which even the glacier ends themselves are at times buried beneath extensive alluvial deposits. Since there is no persistent advance of the piedmont bulbs there is no opportunity for the development of fringing moraines. The glaciers attain a position during advance and spreading, and they retain this position for a long time, not through continual supply, but because of the protection from ablation which their moraine cover supplies. Abundant water issues during the long period of wastage, and it bears enormous loads of sediment. Hence water deposits assume immense importance. Probably far the greater part is deposited in the neighboring sea, to which nearly all of the clay is carried; but the greater portion of the bowlders, pebbles, and sand accumulate either in alluvial fans fringing the glaciers, in deltas on the fiord shores, or in the beaches between the deltas.

If one of these glaciers should completely disappear, there would be left, as the most conspicuous deposit, a broad, crescentic frontal zone of coalescing alluvial fans, very coarse in texture near the glacier fronts, and grading to gravel and sand away from it, with clay beyond on the fiord bottom. On the inner margin, in favorable places where the gravels rested on ice, a hummocky deposit would develop as a result of irregular settling during the melting of the buried ice, with pond and small lake areas within it. Thus a morainic topography, kame-moraine would perhaps be better—would develop in a more or less perfectly crescentic area around the outer margin of the piedmont bulb, grading outward into the more even slopes of alluvial fans and outwash gravel plains. On the inner side of the morainic crescent would be a depression, with irregular surface and with a veneer of coarse angular fragments. The site of the depression is often occupied by a lake.

Leading from the crescentic moraine, on either end of the crescent, there would extend a marginal band of deposit, also water-laid, marking the sites of the marginal streams, and extending well up into the mountain valleys. These marginal stream deposits

would be exceedingly complex, gravel in the main, but with local lake sediment areas, with patches of moraine, with included plant remains and peat bogs, with alternate layers of two or more of these, and, in places, with gradations to marginal valleys and even rock gorges, where the lateral streams had been forced to cut across rock spurs. All these varieties of conditions are present along the margins of the existing glaciers, and since they have been fully described and discussed elsewhere,[1] they will not be further considered here.

Origin and Effects of Icebergs. Origin and Nature. Vast quantities of ice are discharged from the three tidal glaciers,—Nunatak, Turner, and Hubbard; and the amount of floating ice in the inlet, particularly in Disenchantment Bay, is very great. The larger portion of this ice is in small pieces, but there are also many large icebergs, some rising 40 to 50 feet out of the water, or even more, and having a length of 200 to 300 feet. Many of the icebergs, especially the smaller, are free from débris, but large numbers bear some moraine, while icebergs black with included débris are by no means uncommon. We estimate that about twenty per cent of the icebergs carry a noticeable amount of morainic material, and at all times thousands of tons of débris are being floated in the waters of Yakutat Bay. Naturally most of it is finer material, but many icebergs are seen in which good-sized bowlders are embedded. Since there is little moraine in the portion of the glaciers above water level, and since the marginal moraine-covered portions of the glaciers discharge few icebergs, we assume from the abundance of débris-charged icebergs that there is much moraine incorporated in the lower layers of the glaciers, and that these débris-charged layers extend a considerable distance above the bottom of the glaciers.

The discharge of icebergs is almost constantly in progress from the front of the largest tidal glacier, the Hubbard, and at frequent intervals from the other two. One need look at the Hubbard front but a few minutes to see the fall of icebergs, while the sound of their discharge, and the waves to which they give rise, are almost incessant. The most noticeable discharge is that from the ice cliff above water level, from which single blocks and avalanches of many blocks are commonly seen cascading down the ice front and sending the spray high in the air. Occasionally great masses tumble down into the fiord, but most of the discharge is in the form of small pieces, a few feet in diameter. Even when a large mass starts, it usually crumbles in its descent and reaches the fiord in small pieces, for the upper ice is evidently not only brittle, but weakened by much breaking. Often, when viewed from a distance, the falls of ice down the glacier front resemble a mass of falling water. These iceberg falls from the tidal cliff are made possible by the attack of the sea water at and below the visible cliff base, by the extensive crevassing which breaks the upper glacier, and by the melting of the glacier in the air, by means of which the disruption of the ice is extended. By these means the rapidly-moving glacier is checked in its advance.

Work Performed. Because of the rapid and abundant discharge of ice from the cliff above tide level a submerged, projecting ice foot is produced from which masses rise every now and then, producing the larger bergs of the inlet, including the glassy and most of the débris-charged icebergs. Probably most of these rising icebergs are small, but both Russell and the senior author have witnessed the rising of large masses, and there are

[1] See Tarr, R. S., The Yakutat Bay Region, Alaska, Professional Paper 64, U. S. Geol. Survey, 1909, pp. 96–106, 120–137; Some Phenomena of the Glacier Margins in the Yakutat Bay Region, Alaska, Zeitschrift für Gletscherkunde, Vol. III, 1908, pp. 81–110; von Engeln, O. D. Zeitschrift für Gletscherkunde, Vol. VI, 1911, pp. 104–150.

many hundreds of large icebergs floating in the fiord at all times, most of which must have come from the lower part of the glacier. When such great masses rise they send out huge waves which cause violent surf on the neighboring coast.

The formation and dispersion of icebergs produces important results in addition to checking the advance of glaciers. One of the most important of these is that of marine erosion resulting from the iceberg waves. Within three or four miles of Hubbard Glacier front iceberg waves are almost constantly breaking, and frequently with such violence that it is sometimes difficult to land, even on a beach, and dangerous to attempt a landing on a rocky coast. These waves are accomplishing far more work of erosion than the wind waves, and they are both building beaches and cutting sea cliffs. Gilbert found, what our observations abundantly confirm, that some of the sea cliffs near the tidal glacier fronts are at a higher level than these waves now reach, evidently having been formed when the glacier fronts were nearer and the waves, therefore, both higher and more vigorous.

Since practically all of the icebergs melt within Yakutat Bay they are important agents in transportation of the sediment which is accumulating there. They contribute both to the beaches and to the sediments away from the shore, which are mainly clay brought by the glacial streams. One may be confident that scattered through these fine-grained sediments is a notable admixture of coarser ice-borne fragments, even to the size of large bowlders. By this iceberg dispersion of glacier-borne material, the rate of upbuilding of moraine deposits at the front of tidal glaciers is greatly diminished; and when one finds such an extensive submarine moraine as that which sweeps in a broad crescent at the mouth of Yakutat Bay he may feel certain that the ice stood there for a long time. Probably, too, it is composed in large part of materials poured into the sea by the glacial streams.

Many of the icebergs are stranded on the beaches. At times the west shore of Disenchantment Bay and Yakutat Bay as far as the Kwik River is so covered by stranded bergs that it is difficult to land a boat, especially when the surf is breaking. These stranded icebergs melt rapidly and contribute much material to the beaches; the surf, swirling among them, moves and grinds up the beach material; the icebergs are rocked and pushed back and forth, aiding directly in erosion; and only rarely do they form such a rampart as to check erosion by completely breaking the force of the waves. Off shore, large bergs are often stranded, and when they run aground, they must plow up the bottom material, while when they break and the fragments rock back and forth, they must cause still further erosion, and the waves to which they give rise add materally to the surf work on the neighboring beaches. Many of the stranded bergs remain for several days in one place and must give rise to local deposits of coarse material, forming pockets in the midst of plowed up and disturbed clay and sand sediments. The same phenomenon is often observed on the beaches where icebergs have stranded at high tide, and on melting, or floating off, have left pockets of angular rock fragments in a pit in the sand, where the iceberg stood. So much coarse material is brought by the icebergs that all the beaches in the zone of abundant bergs contain a notable admixture of coarse fragments, including bowlders of good size, especially at and near the low tide mark, where large icebergs often strand.

Dispersion of the Icebergs. The dispersion of the icebergs of Yakutat Bay is very peculiar, for, notwithstanding their abundance and the large size of many of them, it

is rarely the case that one is seen in the broad outer part of the bay, and so few escape to the ocean that they are not considered a menace to navigation. A broad and constant stream of icebergs passes out of Disenchantment Bay, hugging the west shore, though broadening somewhat in the outer portion; but beyond the Kwik River few are ever seen, and they are normally rare in the central and eastern parts of the bay. Practically the entire mass of ice discharged by the three tidal glaciers is melted either in the waters of the inlet, or by stranding on the west shore north of the Kwik River. This is very different from the condition at Columbia Glacier where the waters of the inlet are not heavily burdened with icebergs, for, in spite of the great amount of ice discharged from the glacier, a few of the bergs are carried out of the fiord into Prince William Sound. In Yakutat Bay some set of conditions prevents the discharge of the icebergs through the broadly-open bay mouth.

The Ice Jam of June, 1910. In June, 1910, the junior author encountered an ice jam in Yakutat Bay which seems to have been caused by a combination of unusual conditions which was not present at the times of our earlier visits. When we sailed into Disenchantment Bay on June 12, 1910, we found much heavier ice at the native sealing camp than in previous years, and from that point to Osier Island the boat was forced through the ice pack with the greatest difficulty. There was much less clear water between the sealing camp and Calahonda valley than usual, and from the south end of Haenke Island to Osier Island there was a solid jam of icebergs without any open lanes whatever. Near shore, as well as out in the bay, the conditions were the same, and only by pushing the bergs aside with boat hooks and oars were we able to keep the bow and propeller sufficiently clear so that the engine could keep the boat progressing. There was an enormous and unusual proportion of bergs so large that they could not be pushed aside, and the danger of their overturning and swamping the boat kept us constantly on the alert. The tidal currents also added to the difficulty by drifting towering icebergs down upon the boat when it was temporarily at a standstill. It took us from 5 in the afternoon to 10.30 at night to travel the last part of the way, a distance which the boat ordinarily would make in less than an hour. The next day we could see the ice jam clearly and found Disenchantment Bay and Russell Fiord tightly packed with icebergs (Pl. XC) from the sealing camp to Marble Point, an area of between 40 and 45 square miles. This is exclusive of a great number of icebergs in outer Yakutat Bay, the west side of which is commonly filled with icebergs, as already stated. The east side also had many more bergs than is common, as is stated later, and there were more than usual in Nunatak Fiord and the south part of Russell Fiord. Between the sealing camp and Marble Point the fiord was absolutely filled with ice, a condition which we have never before observed.

On the evening of June 13th there seemed to be a narrow strip of clear water on the north side of Russell Fiord southeast of the delta of Variegated Glacier; but on the 14th, when we tried to cross to this from Osier Island the ice jam was still so compact that the launch was unable to make headway against it. We struggled for six hours (from six in the morning to noon), and travelled less than a half mile. The smaller ice fragments had frozen together the night before, some of the conglomerate cakes being strong enough to bear the weight of a man, and it was necessary to break these apart with an oar or axe before the boat could penetrate through them. There were hundreds of very large bergs where in previous years we had commonly seen a dozen or less, and between them was such a solid pack of smaller fragments that no water was visible in the fiord. It was

necessary to turn the boat this way and that, and four of us were busy all the time breaking ice, pushing bergs from under the bow and away from the propeller while the other three respectively steered the boat, started, stopped and backed the engine, and attended to the smaller boat we were towing. Sometimes we were stopped absolutely for several minutes at a time and now and then the tide drifted the largest icebergs, which we attempted to avoid, down close to us. When one of these bergs, rising fully 100 feet above water, and therefore six to seven hundred feet from base to crest, turned over and went to pieces a hundred yards from us we were not a little dismayed, for if the same thing had happened when we were a little nearer it would have been entirely impossible to go away or even to turn the boat bow-on to the gigantic wave which spread from the crumbling iceberg. The filling of the boat by a wave, or the falling of one of the ice fragments into the launch, would have resulted in instant disaster. These personal adventures are narrated here to show the dense character of the ice jam on June 12th, 13th, 14th and 15th. Venturesome as some of the party were, it was agreed by all that the mass of icebergs was not to be penetrated by our boat, and after struggling to the lee of a reef and anchoring at noon in the hope of better conditions after the tide had turned, we were temporarily forced to give up the struggle with the ice.

On June 16th, however, there was less ice in Russell Fiord than on the four previous days, and we made our way slowly through the ice pack to Marble Point, above which the fiord had many clear lanes.

During the next two days when we were busy in Nunatak Fiord, Seal Bay, and the head of Russell Fiord nothing was seen of the conditions near Osier Island, and on our return June 19th the ice jam had cleared to such an extent that we were able to find enough clear lanes so that an easier trip was made back to the vicinity of the native sealing camp, in the course of which a landing was effected on the border of Hubbard and Variegated Glaciers and a few soundings were made between Osier and Haenke Islands and the Hubbard Glacier. The waves from falling ice pinnacles on Hubbard Glacier were damped by the blanket of icebergs covering the fiord so that there was little danger from them while we were in the thicker ice.

On June 20th, and again the next day, we attempted to force the launch through the ice of southern Disenchantment Bay to the west shore, where we were most anxious to visit Lucia Glacier. Similar conditions to those near Osier Island were encountered, and although we went south into outer Yakutat Bay some distance toward the Kwik River we were unable to cross the bay. The huge bulk of giant icebergs looming through a dense fog added to the difficulty here and it was impossible to see any distance ahead. We steered by compass, though with innumerable detours, and were eventually turned back because of the entire absence of open lanes through the ice. On June 22nd it was also impossible to cross Yakutat Bay to the west side because of the ice jam and on this and the preceding days we forced our boat through much thicker ice than we had penetrated in the previous years' work in Yakutat Bay, making a few soundings along the east shore of Disenchantment Bay. Going down to Knight Island, very heavy ice was encountered in the broad outer part of Yakutat Bay, the pack filling the bay to the east shore some distance south of Logan Beach. Except for scattered fragments we have never before seen floating ice on this east shore. Residents at Yakutat Bay told us that many icebergs came ashore earlier in June as far south as Khantaak Island.

Summarized, therefore, the ice jam of June, 1910, in Yakutat Bay, Disenchantment

Bay, and Russell Fiord was a feature not observed in previous years. It was heavier near Osier Island on June 12th to 16th than on the 19th and near the sealing camp on June 19th to 22nd than on the 12th. We thought of it at first as possibly due to a great advance and increase of iceberg discharge from Hubbard or Nunatak Glacier, or both. This was perhaps a contributing cause, but the clearing up of the ice jam, even while we were there, shows that it was not the chief reason. As a matter of fact neither glacier had advanced much, and Nunatak Glacier which moved out the farthest of the two did not supply Nunatak Fiord with many more icebergs on June 17th than during the previous years.

The ice jam seems to be due mainly to a climatic combination. There was an exceptionally heavy snowfall in 1909–10 and a late spring, for photographs show that there was much more snow at sea level in Yakutat Bay on June 11–26, 1910, than at the time of Russell's visit on July 3, 1890, or that of the Harriman Expedition on June 19–22, 1899. There was far more than during our previous visits. The local climate was therefore colder, icebergs melted more slowly, and their freezing together at night hindered free floating with the tide. These conditions, combined with a little increase of iceberg discharge, are doubtless responsible for the ice jam of June, 1910. Such a jam may occur, in less accentuated form, every spring. Our previous visits have perhaps been just too late to witness it, though natives have reported heavy ice which interfered with sealing operations just before our arrival in several previous years. There was probably less ice just before our visit in 1910, for a bear hunter and the missionary of Yakutat had each penetrated into Russell Fiord in a smaller and less powerful boat than ours not long before our visit in the middle of June. They encountered less ice than we did.

Since the ice jam formed after these visits and cleared up somewhat while we were there, probably moving outward down the bay, it is thought to be an abnormal occurrence due to such a combination of circumstances as is outlined above.

There seems to have been a similar ice jam in Yakutat Bay in the spring of 1911, as is indicated by the following letter.

Branch Hydrographic Office,
Port Townsend, Wash.
April 17, 1911.

NOTICE TO MARINERS

Captain McGillivray of the Am. S. S. "*BERTHA*" reports that on April 10, 1911 in Lat. 59° 38′, N. Long., 141° 30′ W. passed through a field of drift ice and large bergs. Ship ran 53 miles E. N. E. with ice on all sides as far as 30 miles off shore. Had to stop the ship on several occasions on account of large bergs floating so close together. Ice was probably blown out of Disenchantment Bay. Large bergs were drifting toward the track of ships passing to the eastward of Cape St. Elias.

A. B. Wyckoff,
Lieut. U. S. N.

The latitude and longitude given is off the western edge of Malaspina Glacier and almost directly south of the new Icy Bay (Chapter III). It is quite possible that these icebergs may have come from there rather than from Disenchantment Bay. As drift ice in sufficient amount to interfere with navigation is exceedingly unusual in this part of the Pacific Ocean the occurrence is well worth attention.

Glacial Sculpture Below Sea Level. Glacial erosion above sea level was studied in our investigation before 1910 and earlier by Russell and Gilbert. Below sea level the form of the bay and fiord are revealed by soundings, which were first carried on systematically

by the junior author in 1910. Earlier soundings had been made as follows. In 1786 the French explorer, LaPerouse, recorded depths of water in the mouth of outer Yakutat Bay. In 1787 the English explorer, Dixon, sounded in Port Mulgrave, the harbor at the east entrance of Yakutat Bay. In 1791 Malaspina, an Italian in the service of the Spanish, made a few soundings in lower Disenchantment Bay between the native sealing camp and Haenke Island, and made a good map of Port Mulgrave with soundings.[1] In 1788, 1793, or 1807 Russian explorers made three or four soundings in western Yakutat Bay near Galiano Glacier.[2] Other soundings were made in Icy Bay near the west side of Malaspina Glacier by these Russians and by several English explorers.

In 1890 the American Revenue Cutter *Corwin* took Russell into Disenchantment Bay, making two soundings, which an officer of the ship has since roughly located for us, near the native sealing camp and west of Haenke Island. They also made several unrecorded soundings, and Russell[3] says that "soundings made between the island (Haenke) and the ice foot (Hubbard) gave forty to sixty fathoms," and that "a few soundings made in Disenchantment Bay within half a mile of the land showed a depth of from 40 to 120 fathoms." In 1891 Russell attempted to determine the depths of water in what we now call Russell Fiord and states[4] that "with a line 170 feet long we could get soundings only in the smaller coves and occasionally within a few rods of shore." In 1892, a century after the early, precisely-located soundings by Malaspina, the U. S. Coast and Geodetic Survey made an excellent chart of outer Yakutat Bay, showing the depths of water from the Pacific to the mouth of Disenchantment Bay. This is the basis for the bathymetric map by Gilbert[5] which is reproduced in this book as Fig. 21. The Harriman Expedition made no recorded soundings in 1899, though they carefully investigated with the lead before taking the *George W. Elder* through Disenchantment Bay, Nunatak Fiord, Seal Bay, and to the very head of Russell Fiord, demonstrating that an ocean-going steamer can sail in all parts of this deep inlet, where floating ice does not prevent. We ourselves made no soundings in 1905 or 1906. In 1905, however, we did locate several new reefs, north of Haenke Island and east of Knight Island, which were uplifted during earthquakes in September, 1899.[6] We also made several hundred measurements of amounts of vertical change in about 120 miles of shoreline, where there were uplifts of from 7 to 47 feet, submergences of from 5 to 7 feet, and unknown changes offshore, where there was faulting along some of the fiords. In 1909 an attempt was made to carry out a systematic series of soundings, but the attempt failed because the apparatus which we had proved to be inadequate.

In 1910, the junior author, ably assisted by Mr. E. F. Bean of the University of Wisconsin, made soundings throughout Russell Fiord and all of Disenchantment Bay except the iceberg-crowded western portion. A boat sounding apparatus loaned by the U. S. Coast and Geodetic Survey made it possible to secure most satisfactory results which are

[1] These French, English, Spanish, and Russian soundings are plotted in the respective reports and atlases and reproduced by Russell in Nat. Geog. Mag., Vol. III, 1891, Plates 3, 4, 6, and 7.

[2] Shown in Tebenkof's Atlas, 1852, and reproduced by Davidson in Trans. and Proc. Geog. Soc. Pacific, 1904, Pl. 6.

[3] Russell, I. C., Nat. Geog. Mag., Vol. III, 1891, pp. 56 and 100.

[4] Russell, I. C., 13th Ann. Rept., U. S. Geol. Survey, Part II, 1892, p. 90.

[5] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, Fig. 27, p. 50.

[6] Tarr, R. S. and Martin, Lawrence, Bull. Geol. Soc. Amer., Vol. 17, 1906, Pl. 23 facing p. 54. These reefs are also referred to in U. S. Coast Pilot, Alaska, Part I, 1908, pp. 185–186.

plotted in Pl. XCI and our contour maps. The depths shown were all obtained in 1910 except the one of 264 feet west of Haenke Island, which was made by the *Corwin* in 1890, two of 630 and 1002 feet respectively in the mouth of outer Yakutat Bay, made by the Coast Survey in 1892, and two of 720 feet (no bottom), south of Haenke Island, made by Malaspina in 1791. None of these soundings in 1910 are corrected for mean sea level and they were taken at various and unrecorded stages of tide. As the extreme range of tides at Yakutat village is seven to ten feet, and the maximum in Russell Fiord probably not greatly in excess of this, the greatest error in any sounding is surely less than ten feet, which for the purposes of this discussion is negligible. Not quite all our soundings are shown on Pl. XCI, because of the small size of the map. Each sounding is plotted so that the middle of the figure is at the location of the sounding.

These soundings show specifically, what we had previously inferred, that the fiord is very deep. It reaches a maximum of 1119 feet below sea level in southern Russell Fiord, where it exceeds even the greatest recorded depth in outer Yakutat Bay. The depths in Disenchantment Bay range from 150 to 942 feet, with possibly greater depths in the western half of the bay, where the ice jam made sounding impossible. There is a channel east of Haenke Island sloping regularly southward from 264 to 576 feet. Northwestern Russell Fiord slopes southeastward from 216 to 888 feet, being shallowest, as Disenchantment Bay is also, near Hubbard and Variegated Glaciers. Nunatak Fiord has depths of from 261 to 666 feet, not sloping regularly as northwestern Russell Fiord does. Southern Russell Fiord has depths of from 786 to 1119 feet, with rapidly shoaling water in Seal Bay near Hidden Glacier and at the head of the bay where the fiord emerges from its mountain walls and expands in the Yakutat Foreland. In all parts of the inlet the water, of course, shallows rapidly near shore but, as Russell observed, the fiord is everywhere deep and steep-sided.

The great depth below sea level, the form of the submerged topography, and the departures from normal slopes, etc., are all explained satisfactorily by glacial erosion, which seems to have completely erased the structural or stream-carved-and-submerged preglacial topography. The features in connection with glacial deposits below sea level will be taken up on a subsequent page. Those due to glacial sculpture fall under the headings of (a) the fiord cross-section, (b) the longitudinal bottom profile of fiords, (c) the submerged hanging valleys.

The fiord cross-section (Figs. 17 and 18) is plainly that of a round-bottomed V, as Davis has phrased it, rather than the often-quoted U-shape. The slopes above and below sea level are not essentially different. Knowing the depths of water in the fiord, and the steep-sided character of the fiords, the bulk of rock eroded in the formation of these fiords could be computed and the result would be a striking figure. How rapidly the water deepens offshore in places is shown by the fact that a sounding 100 yards from the coast near Pt. Latouche at the east entrance to Disenchantment Bay gave 312 feet, a slope of over 45 degrees. The fiord cross-section is a simple one and its large pattern is that of glacial erosion, in contrast with the angular pattern of faulting or the complex, small-textured pattern of stream erosion.

The longitudinal bottom profile of fiords, brings out three points: (1) we cannot distinguish irregularities due to glacial erosion from those due to glacial deposition below water by soundings alone; (2) the vertical range in the bottom profile constitutes a marked contrast between fiord-bottom profiles and those of stream valleys; (3) because of the

flattish bottom revealed in cross-sections and the straightness of fiord walls, a number of similar, parallel, bottom profiles exist in every fiord.

The first point is well brought out by comparing the submerged contours of northwestern Russell Fiord with those of Nunatak Fiord and southern Russell Fiord. The bottom profile of the former reveals a persistent southeast slope at the rate of nearly sixty feet to the mile, while the latter do not slope consistently but have marked up and down grades. These might perfectly well be due either to irregular glacial scooping or to moraines below sea level, resting on a simple slope similar to that in northwestern Russell Fiord. With glaciers of this size, differential glacial erosion of two or three hundred feet in a distance of several miles in a longitudinal valley is perfectly possible, as glacial rock basins in various regions testify, but submerged moraines of this height are also possible. It is enough to point out here that the bottom profiles are thus interrupted, as in Nunatak Fiord and lower Russell Fiord, in contrast with regular, uninterrupted, bottom slopes in northwestern Russell Fiord and the channel east of Haenke Island in Disenchantment Bay, and that we ascribe some such interruptions in grade to glacial scooping, and others to submerged moraines. Imperfectly eroded knobs surrounded by

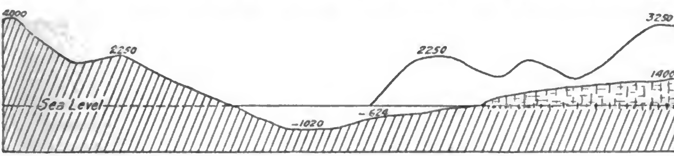


FIG. 17. SECTION OF SEAL BAY AND HIDDEN GLACIER WITH ITS SUBMERGED HANGING VALLEY. CROSS-SECTION OF RUSSELL FIORD. VERTICAL AND HORIZONTAL SCALES THE SAME.

deep water also exist, such as the 12 foot shoal near Osier Island where the *Princess Maquinna* grounded in 1913.

The second point, regarding longitudinal fiord bottom profiles in contrast with longitudinal profiles of stream valleys, cannot be discussed adequately with the existing doubt regarding the cause of these interruptions of fiord bottom slopes. If some of the two or three hundred foot interruptions in slope are in rock, and not due to moraine accumulations, we have an excellent argument in favor of the glacial origin of these fiords, for upgrades of 200 to 300 feet are impossible in stream valley profiles, whereas rock basins would naturally develop in fiord bottoms or in valleys above sea level when eroded by ice.

The map shows clearly that many parallel longitudinal bottom profiles are possible with the width of the flattish fiord floors, as in northwestern Russell Fiord. This contrasts markedly with the bottom of a submerged, stream-eroded valley where the medial bottom profile is rarely straight and almost never could have a single similar profile parallel to it. There are no submerged spurs entering the fiord from either side. The bottom profiles, longitudinally, show that the fiord is exactly similar in plan above and below water and is characteristically of the large, simple pattern produced by great ice erosion.

The submerged hanging valley is well illustrated by Seal Bay (Fig. 17) which is from 180 to 600 feet deep and hangs approximately 400 feet above Russell Fiord. We think of no essential differences in origin between the hanging valley above and below sea

level and in this case it is very clear that the four hundred feet of discordance measures the difference in erosive power of the former Russell Fiord Glacier and the smaller ice stream of its tributary, the Hidden Glacier. That the Hidden Glacier was not wanting in erosive power is evidenced by its steep valley walls below sea level in Seal Bay as well as above sea level near the present glacier. The lip of this hanging valley also testifies, by its position well out in the fiord, that the Hidden Glacier pushed the Russell Fiord Glacier well to the west at their junction, thus lessening the discordance of junction. That Hidden Glacier did not erode as fast as the main ice stream, however, is shown conclusively by this submerged hanging valley.

The hanging valley condition also exists near Cape Enchantment, the junction of Nunatak and Russell Fiords, where the distributary of Hubbard-Variegated Glacier moving southeastward in the northwest arm of Russell Fiord was evidently stronger than the former Nunatak Glacier. The amount of discordance may be as great as 444 feet or as little as 200 feet, depending on whether there is a submerged moraine here. It is clear from these soundings near Cape Enchantment, and from the upslope toward Varie-

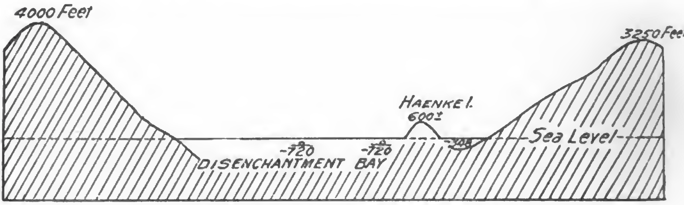


Fig. 18. Cross-Section of Disenchantment Bay. Vertical and Horizontal Scales the Same.

gated Glacier, that the ice from Nunatak Fiord did not move northwestward toward Hubbard Glacier for a very long time.[1] That this hanging valley is not due to faulting is shown specifically by two soundings of 888 feet, on either side of a fault line which traverses the fiord bottom. The cliff of the hanging valley is a mile east of this fault.

A third pronounced case of topographic discordance below sea level is the channel east of Haenke Island in Disenchantment Bay, whose relationship to Disenchantment Bay is very much like that of the valleys north and south of the nunatak of Nunatak Glacier. The end of this channel hangs at least two or three hundred feet above the main Disenchantment Bay, the depths of water in the western part of which are still partly undetermined.

It is to be noted that these submerged discordances (200 to 450 feet) are less than those above sea level, for example 1300 feet in one hanging valley of Nunatak Fiord, and perhaps 1500 or 2000 feet in others. This is because glaciers which are large enough to descend to sea level usually are less discordant in size and power than those small glaciers which do not reach the sea.

The soundings also reveal the total amount of discordance of a number of hanging val-

[1] That one of the last movements was northwestward is amply proved by descending glacial grooves, roche moutonnées forms with crag and tail details, etc. See Professional Paper 64, U. S. Geol. Survey, 1909, p. 119.

leys whose mouths are not submerged. Calahonda valley, for example, hangs about 600 feet above eastern Disenchantment Bay, minus the slight amount to which its rock lip, now buried by outwash gravels, lies below sea level. The Aquadulce valley, at the native sealing camp, probably hangs more than 900 feet. Several small valleys on the west side of lower Russell Fiord hang one or two hundred feet above sea level, but their total discordance to the bottom of Russell Fiord is 1000 or 1200 feet. The visible 700 foot discordance of a prominent hanging valley in Nunatak Fiord is increased to over 1300 feet by the depth of 612 to 666 feet of water in the fiord.

It is impossible to state how much Turner Glacier hangs above the bottom of Disenchantment Bay, for soundings are lacking here. Turner is much smaller than Hubbard Glacier and the discordance should be great. The visible discordance is at least four or five hundred feet, to which the nearest sounding adds only 276 feet. Between this sounding and the Turner Glacier, however, is just where the powerful Hubbard Glacier should have eroded most deeply within the steep walls of Disenchantment Bay, and we should be surprised if the water is not over 900 or 1000 feet deep in that part of the fiord.

Submerged Glacial Deposits. The soundings in the fiords tributary to Yakutat Bay add to our knowledge of the glacial deposits below sea level. It is unfortunate that time was not available for determining the material on the bottom of the fiord and for taking

FIG. 19. LONGITUDINAL SECTION OF NUNATAK FIORD AND GLACIER. VERTICAL AND HORIZONTAL SCALES THE SAME.

temperatures, etc., and, in the absence of bottom samples, little can be said of the specific nature of the submerged glacial deposits. We can speak more definitely, however, of the topography of some of the underwater deposits.

In Nunatak Fiord the soundings reveal a shoal a little less than 1½ miles from the 1910 ice front (Fig. 19). Upon this the water is 261 feet deep in mid fiord so that the shoal rises 294 feet above the fiord bottom to the east (toward the glacier) and about the same amount above the more distant deep region to the west. This may be either a moraine or a rock ridge. The ice front stood here as recently as 1899 but we do not know for how long. We had previously suspected a shoal here, for large icebergs habitually stranded upon it, especially near the north shore, in 1905 and 1909. It may be morainic, as Fig. 19 suggests. It is of undetermined width and descends most steeply on the east side (suggesting an ice contact), where the slope is not less than 300 feet to the mile, and probably much less steeply on the west where the slope is between 100 and 200 feet to the mile and flattens to 84 feet to the mile and less in a short distance where there may be stratified deposits built by former streams from the ice and by fiord currents. As the glacier is known not to have halted very long at this 1899 position we regard it as possible that the submerged moraine is built upon a slight irregularity of the fiord bottom, due to differential glacial erosion; but we do not know enough of the relative resistance to glacial erosion by the rocks making the fiord bottom to determine this, and it is considered possible rather than probable.

A second shoal which may be wholly or in part a submarine moraine is at the junction

of Disenchantment Bay and Russell Fiord, near Osier Island (Fig. 20) where the water is only 216 to 339 feet deep in contrast with depths of 500 to 800 feet near by in Russell Fiord and probable depths of 900 to 1000 feet in Disenchantment Bay. This may be regarded as a moraine rather than a place where the fiord was originally shallow (1) because of the suggestive slope of the mountain wall behind Osier Island, which should presumably descend as steeply below sea level as it does above, judging by other cross-sections of the fiord, and (2) because Hubbard and Variegated Glaciers are directly behind the shallow region and are still building up a deposit near by. This is being built up by direct discharge of the débris carried to the ice front of Hubbard Glacier, and by sediment from the streams beneath and at the margin of Hubbard, and from the Variegated and Orange Glacier streams whose great volume, muddiness, and visible deltas show how much detritus is being carried into the fiord. Beyond the ice front and the shore the floating icebergs are also adding to the submarine deposits. The fiord is very shallow, certainly, as might be natural at the bifurcation of this valley where erosion is not concentrated. Unfortunately we do not know the depths in the uncharted waters of the west side of Hubbard Glacier and in front of the Haenke and Turner Glaciers. The moraine, if such it be, is somewhat abnormal in plan because built at the junction of two fiords and its surface shows nothing of the steep ice-contact feature of the

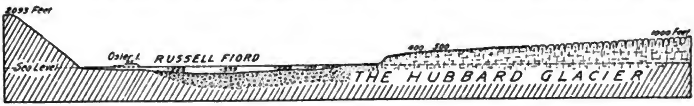


Fig. 20. Cross-Section of Russell Fiord at Hubbard Glacier. Vertical and Horizontal Scales the Same.

supposed moraine near Nunatak Glacier, because the ice front has not yet retreated from it. The water is deepest in mid-fiord, as might be expected, but within a hundred yards of Variegated Glacier the water is 105 feet deep and increases rapidly westward. We do not regard it as demonstrated that this is a moraine rather than an originally shallow part of the fiord.

A third shoal, which pretty surely represents a moraine under water, lies about 7½ miles north of the head of Russell Fiord and is of especial interest because we had previously postulated a halt of the ice front of the expanded Russell Fiord Glacier at about this point, in connection with (a) overridden glacial gravels (b) the beaches of a marginal lake in the lower end of Russell Fiord.[1] Both Russell and Gilbert recognized the evidence of this former lake[2], and Gilbert detected the signs of overriding in the gravels.[3] There should be a moraine at the junction of the southward limit of overridden gravels and the northward limit of the abandoned beach which were first shown upon a map in 1905; and just here the submerged moraine was found in 1910. The water is 828 feet deep at this point and 1119 and 1044 feet on either side, so that the height of the submerged

[1] Tarr, R. S. and Martin, Lawrence, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 163–4 and map facing p. 145; Tarr, R. S., Professional Paper 64, U. S. Geol. Survey, 1909, pp. 132–134, Pl. XXXVII and Fig. 9.

[2] Russell, I. C., 13th Ann. Rept., U. S. Geol. Survey, 1892, pp. 88–89; Amer. Journ. Sci., 2nd Series, Vol. XLIII, 1892, p. 173.

[3] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 48, 51.

moraine above fiord bottom is at least 281 feet on one side and 216 on the other. The inferred conditions in southern Russell Fiord before, during, and since the building of this submerged moraine are shown in the three maps of Pl. XCII. If this is wholly a moraine the ice must have halted here a long time, for accumulation under water must be very slow indeed.

In outer Yakutat Bay the great shoal at the entrance is surely morainic, and there may be a second moraine below sea level northwest of Knight Island (Fig. 21) as Gilbert has shown.[1] It is of interest to compare this outermost moraine of the Malaspina-Disenchantment Bay Glacier with the outermost moraine of the Russell Fiord Glacier. The former (Fig. 21) is two or three miles wide, rising 300 to 400 feet above the bottom of the inner bay, and sloping steeply seaward on the Pacific side, where there is no marked convexity. This contrasts with the Russell Fiord terminal moraine whose inner face is revealed by the soundings in 1910 (Pl. XCII). The Russell Fiord moraine is nearly twice as wide, rising to approximately 200 feet above sea level, or 275 to 980 feet above the bottom of the expanded end of Russell Fiord. It is markedly convex seaward and is bordered by a gently-sloping outwash plain. Because the glacier was not cut back by the ocean waves and melted back by the salt water it had the form of a bulb glacier.

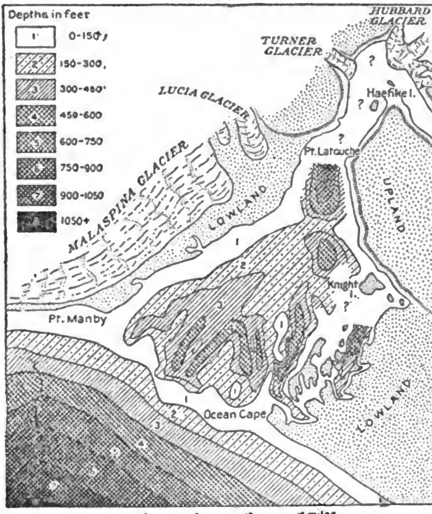


FIG. 21. YAKUTAT BAY, SHOWING CONTOURS OF DEPTH AND SUBMARINE MORAINE BETWEEN OCEAN CAPE AND POINT MANBY (AFTER G. K. GILBERT).

Other topographic features suggestive of either moraines below sea level or ridges between rock rimmed basins of the fiord bottom (Pl. XCI) are (1) near Shelter Cove in southern Russell Fiord; (2) between Seal Bay and Cape Enchantment; and (3) immediately east of the lip of the hanging valley of Nunatak Fiord. In each case the topographic form of the fiord bottom indicates either a morainic shoal or an irregularity in glacial sculpture. The first of these shoals rises 75 to 100 feet above the general fiord

[1] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 49–50.

bottom and is in a narrow place in the fiord, though the mountain walls are the normal distance apart. Here the lower part of the glacier should erode most effectively instead of leaving a shoal. On the other hand, there are no abandoned beaches extending up to this place as in the case of the submerged moraine three miles south of it and we are not inclined to think that the irregularity represents a halt of the ice.

In the second case the upgrade of 210 feet in the southward sloping fiord bottom south of Cape Enchantment may be due to imperfect glacial sculpture of a massive conglomerate [1] which outcrops just here on the fiord walls. The third irregularity, at the mouth of Nunatak Fiord, rising 200 feet above the fiord bottom to the east, is also possibly explained either by glacial sculpture or as a submerged moraine, between which it is impossible to discriminate with the facts available.

The soundings made in 1910 also establish the fact that, deep as the water is, it is practically impossible that any of the glacier fronts of Disenchantment Bay and Russell Fiord are floating now and they do not seem to have been afloat at any stage of their expansion, judging by the depths of water. This means that there was always active glacial grinding on the fiord bottom and the problem arises as to where this eroded material is now. In Russell Fiord the volume, merely from the part of the fiord below sea level would be many cubic miles, and the soundings show that more material was eroded above sea level than below. Some of this material makes up the great moraine south of Russell Fiord, some is in the submerged moraines, and a great deal has gone out to sea. Some, however, doubtless remains in the fiord bottoms, making it impossible to tell how near a given sounding goes to the rock bottom of the fiord. The measures of glacial erosion are, therefore, all minima.

In Disenchantment Bay and Russell Fiord we have no determinations of bottom material, but in outer Yakutat Bay the Coast Survey chart [2] supplies some data, though not enough for a thorough study. It is evident that the sedimentary material would be chiefly clay in depths of water such as this, and the offshore slopes are such that sand should not normally be carried very far offshore. In Disenchantment Bay a section sounded in 1910 off the delta of Calahonda valley was as follows:

Distance from shore	*Depth*
1-8 mile	12 feet
3-16 "	111 "
1-4 "	294 "
1-2 "	672 "

The initial slope of the submerged delta end is at the rate of 1000 to 1200 feet to the mile, about the same as the rock surface upon which it lies.

A similar section off the delta of the Hidden Glacier stream in Russell Fiord shows a first slope at the rate of over 1500 feet to the mile, which soon flattens out.

Distance from Shore	*Depth*
About 100 yards	6 feet
1-8 mile	195 "
1-4 "	180 "

[1] Tarr, R. S. and Butler, B. S., Professional Paper 64, U. S. Geol. Survey, 1909, Pl. XXXVII.
[2] Chart 8455, U. S. Coast and Geodetic Survey, 1901 and 1910.

Distance from Shore	*Depth*
3–8 mile	222 feet
1–2 "	216 "
5–8 "	321 "
3–4 "	390 "

Because of the steep-sided character of the fiords in these sections, the deltas slope so steeply into deep water that sand should not be found very far offshore.

In outer Yakutat Bay, where we have both soundings and bottom determinations, the Coast Survey chart shows sand in depths of 27 to 30 feet off the borders of the big alluvial fans of Malaspina Glacier. The designations "rocky" and "small stones" are used in certain places near the mouth of the bay, from which we infer there are submerged moraines. Clay is the material over nearly the whole of the bottom of outer Yakutat Bay, the chart showing blue mud and black mud in a great many soundings in depths of from 120 to 630 feet. There is far more sand near the west than near the east shore, where there are now no great streams adding to the submerged glacial deposits.

Three things stand out in connection with the submarine deposits in Yakutat Bay and they are all related to the glacial character of these sediments. (1) These muds and clays, when consolidated, will make shales of considerable area and great thickness, and there will be far more shale than sandstone; (2) the shales will be largely unoxidized, because so large a proportion of the material was ground by the glacier from unweathered rocks; (3) they will be distinguished from normal shales and may be identified as marine, glacial deposits by the angular, striated pebbles and bowlders scattered through them by floating icebergs. Enormous deposits of this sort are accumulating in Yakutat Bay and its branches, Disenchantment Bay and Russell and Nunatak Fiords, especially near the ice fronts of the Turner, Hubbard, and Nunatak Glaciers and also where the streams from Variegated, Hidden, Butler, Fourth, and smaller glaciers, and Black, Galiano, Atrevida, Lucia, and Malaspina Glaciers are pouring out sediments for submarine glacial deposits. These far exceed in extent the till and bowlder ridges, which rise here and there above the fiord bottom.

Contributions to Interpretation of Glacial Phenomena

The Yakutat Bay region has given us a number of facts of observation which have a direct bearing upon the interpretation of glacial phenomena in regions of former glaciation. It is a region of great existing glaciers, a region from which still greater glaciers have only recently retreated, and a region of cool temperate climate, which has given rise to phenomena more clearly resembling those of the wasting glaciers of the Glacial Period than are observed in the margins of the larger glaciers of the colder frigid zones. To the student of glacial phenomena most of the applications to interpretation of glacial features in other regions will be apparent from a reading of the preceding chapters; but the matter seems of sufficient interest to warrant its specific consideration in a few brief paragraphs.

The Piedmont Condition. Where glaciers descended from mountain valleys upon bordering plains or plateaus, or into broader valleys, they expanded during the Glacial Period, as they do now in Alaska. Piedmont bulbs and piedmont glaciers spread out at the base of the mountains in Switzerland, Germany, Austria, and Italy; they devel-

oped in western North America, and doubtless in all mountain areas where the conditions resemble those of Alaska. Any student of the phenomena to which these ancient expanded glaciers gave rise receives aid from the results of the study of the living Alaskan representatives of the type.

Deposits of Piedmont Glaciers. The extensive crescentic moraines of the Alpine piedmont glaciers, the extensive development of associated outwash gravel deposits, and the depressions within the crescentic moraines, often occupied by lakes, are so like the Alaskan conditions as to make it certain that their interpretation will be aided by a knowledge of the Alaskan phenomena. This fact was vividly brought to the attention of the senior author in the eastern Alps when, in company with Professor Hans Crammer, he looked out over the site of the great piedmont bulb of the ancient Salzburg Glacier; and later when crossing the moraines of similar bulbs near Munich. In the one case we have the process in operation before our eyes; in the other the finished product; but as Penck and Brückner, Crammer, and others have observed, the nature of the phenomenon is so clear in the finished product that one cannot misinterpret it. In the details, however, light must be thrown on the interpretation of these records of ancient piedmont bulbs from the facts of observations obtained from a study of the living instances.

Glacial Erosion. From Russell's [1] and Gilbert's [2] studies, and later from our own[3] the Alaskan field has yielded facts of importance on the subject of glacial erosion, applicable to other regions of glaciation. Vast erosion has been performed, and many forms resulting from glacial sculpturing have been observed. This important subject is not specifically treated in this book, however, except incidentally in the description of certain Yakutat Bay glaciers and at some length for glacial erosion below sea level as revealed by new soundings in Yakutat Bay in 1910, discussed in an earlier part of this chapter, and in Part III, dealing with Prince William Sound and Copper River.

The features of glacial erosion are so clearly and intensively developed in Yakutat Bay and other Alaskan fiords that applications, comparisons, contrasts, and interpretations in other regions are made easier by reason of the perfection and recency of development of the forms of glacial erosion in the Alaskan region.

Ineffectiveness of Weak Glaciers. In the Yakutat Bay region we have had three stages of glaciers (1) the present, dwindling glaciers, (2) former, greatly expanded ice streams, (3) an intermediate stage of rather weak expansion. The three stages clearly show differences in extent of glacial erosion. Profound work was performed by the greatly expanded glaciers, deepening valleys one or two thousand feet by rock excavation; the less powerful, briefer period of advance failed to remove even the unconsolidated gravels that the glaciers overrode. This aids in understanding the ineffectiveness of small dwindling glaciers of the present day. As Andrews has clearly shown we must distinguish between the ice flood and the present day glaciers, as we must between the stream flood and the low water stage of erosion. The phenomena of glacial erosion in Yakutat Bay throw light on these differences. In the minor erosion of the gravels, by overriding, we have

[1] Russell, I. C., Nat. Geog. Mag., Vol. III, 1891, pp. 100, 191; Amer. Journ. Sci., 3rd Series, Vol. 43, 1892, p. 173; 13th Ann. Rept., U. S. Geol. Survey, Part II, 1892, pp. 85, 90.

[2] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 57, 113, 170.

[3] Tarr, R. S. and Martin, Lawrence, Bull. Amer. Geog. Soc., Vol. XXXVIII, 1906, pp. 158–160; Tarr, R. S., Popular Science Monthly, Vol. 70, 1907, pp. 99–119; Scottish Geog. Mag., Vol. 24, 1908, pp. 575–587; Professional Paper 64, U. S. Geol. Survey, 1909, pp. 107–119.

evidence of the nature of the phenomena that result from such overriding, applicable to other regions.

Vegetation in Glacial Deposits. The observed growth of vegetation at the margin of glaciers, and even on them, and its incorporation in marginal deposits during an advance, point a clear warning against too free use of this class of evidence in the interpretation of the complexity of ancient glaciation. The phenomena of Yakutat Bay glaciation prove conclusively that soil beds, peat beds, and forest trees do not of themselves contribute evidence of interglacial conditions. They may represent no more than a half century or century of stagnation followed by moderate advance. This does not disprove interglacial conditions, nor argue against complexity of the Glacial Period, but it does discredit some of the evidence upon which some of the inferred complexity is based. More than incorporated vegetation is necessary for proof of such complexity.

Origin of Eskers. Few eskers were observed in the Yakutat Bay region, and these were short ones on overridden gravels in Russell Fiord. The absence of eskers is believed to be due in the main to their burial beneath the fringing alluvial fans whose apexes are at the ice tunnel from which the esker-building stream emerges, and which follow up the ice as it recedes. Esker-building conditions apparently exist in the piedmont bulbs, for the streams maintain their courses for a long time.

Our observations prove three important points with respect to eskers (Pl. LXXXIX, A). (1) As already stated, that subsequent alluvial fan deposit may bury previously made eskers. (2) That, even under such favorable conditions as exist in the broad, stagnant, moraine-covered glaciers of this region, eskers are not formed on the glacier surface, for there are no long streams. (3) That moving glaciers are unfavorable to esker formation. The advance of Marvine and Variegated Glaciers completely destroyed pre-existing subglacial drainage; and the point of emergence of glacial streams from such a normally-moving glacier as the Nunatak is frequently changed. From the evidence of the Yakutat Bay region we believe that the eskers are subglacial deposits mainly, if not entirely, associated with stagnant ice conditions.

Outwash Gravels. The extensive sheets of gravel and the rapidity of their deposit in such valleys as that of Hidden Glacier, the absence of growth of vegetation on the outwash gravel plains, and the perfect rounding of the pebbles, are phenomena which throw light on the extent and depth of similar deposits in regions of former glaciation. From a study of such an outwash gravel plain as that which existed in the Hidden Glacier valley in 1899, 1905, and 1906, or that of the Kwik River valley, we can observe the processes of accumulation which undoubtedly existed in thousands of streams issuing from the ancient continental glaciers.

Pitted Plains. Where these gravels rest on buried ice, as in the Hidden and Fourth Glacier valleys, the phenomenon of kettle formation and the production of pitted plains is observed in progress of development, and an explanation of thousands of similar cases is found.

Kames. Where gravels resting on ice settle very irregularly, we see in progress one of the important causes for the formation of kames; and this process is again and again illustrated in the Yakutat Bay region.

Buried Ice Blocks. Perhaps the most important contribution to an interpretation of glacial phenomena in regions of former glaciation yielded by the Yakutat Bay region is the clear evidence of the great importance of stagnant buried ice in shaping the topog-

raphy of glacial deposits. Not merely are kettle and kame areas caused, but extensive depressions also result. The melting of Galiano Glacier, if further deposit should cease, would give rise to a depression of two or three square miles. Such buried blocks may last long, and receive a heavy burden of deposit, and their ultimate melting would give rise to an irregular topography, varying according to conditions, and depressions for lakes or ponds, with irregular bottoms and margins. In regions of former glaciation a multitude of kettles and pond and lake depressions exist whose interpretation is made clear by the assumption of the former presence of buried ice blocks. In our own work in such formerly-glaciated regions we have hitherto often hesitated to assume this hypothetical cause in the absence of direct proof; but with the evidence of the frequency and effectiveness of the buried ice blocks in the Alaskan region there is little reason for such hesitation. We are convinced that on the margin of waning ice sheets in a region of irregular topography, such as prevailed over extensive areas in Europe and America in the closing stages of the Glacial Period, the conditions favoring the burial of detached or partially detached ice blocks, and of stagnant glacier termini, were common, and that the results of their subsequent melting are of great importance in determination of topographic detail.

Summary of the Glacial History of the Yakutat Bay Region

The mountain history of the Alaskan coast region has been long and complex, with several periods of mountain growth and intervals of erosion, and with periods of volcanic activity. The details of this history are not yet worked out, and we have no knowledge as to the climatic conditions. During the Tertiary, and perhaps in the Pliocene division of the Tertiary, coal beds were accumulated, and associated with them are found plants of a warmer or wetter climate. From the topography of the foothills near the coast we infer that there was a long period of subaerial erosion, during which the mountains were greatly reduced and a system of broad valleys developed. This period of erosion, at least in its later portion, was presumably free from extensive glaciation. The mountains are still rising, as the frequent earthquakes testify, and in 1899 there was further notable uplift, in one section of the coast amounting to forty-seven feet. It is probable that the mountains owe their present great height mainly to geologically-recent uplift and that the extensive glaciation of the region is due to the rising of the mountains in recent times. Whether the extension of glaciation accompanied the uplift as Russell inferred, or whether it was preceded by a period of stream erosion in lofty mountains not yet glaciated is not clear; but we are inclined to the latter view because of the extensive series of branching valleys, presumably formed originally by rivers and now occupied by glaciers. All this early history is obscure and at present can be the subject only of hypothesis.

Without being able to assign a period to the beginning of glaciation, we know that the mountains ultimately became clothed in snowfields and the valleys filled with glaciers, and that the condition has lasted a very long time. Whether the glaciers came solely as a result of the rising of the mountains, or whether they were dependent upon a climatic change, giving heavy snowfall to a region already lofty, cannot be positively stated. Nor do we know the history of the initial spread of the glaciers, of what episodes of advance and recession, if any, accompanied it. Ultimately, however, the expansion of the glaciers extended so far as to push the glacier termini beyond the mountain front, filling the entire inlet with a great glacier flood, and pushing the glacier fronts out to the mouth of

Yakutat Bay. That this great expansion of the glaciers was due to some climatic change rather than solely to mountain uplift is indicated by the three following facts (1) the earlier genial climate, (2) the recent and present day recession (both 1 and 2 indicating that the climate of this region has changed), and (3) the widespread extension of glaciers on the northwestern coast from Washington to the Aleutian Islands. While we cannot assert contemporaneity for the advance of glaciation throughout this region, the evidence points strongly toward the conclusion that the expanded stage of glaciation was contemporaneous throughout the region. It seems hardly probable that mountain uplift throughout a distance of some 2000 miles would alone suffice to give rise to such extensive glaciation, and since there has subsequently come an amelioration of the climate, while at a still earlier age there was an even milder climate than now, it is a reasonable inference to draw that the great expansion of glaciers was in response to actual climatic change.

It cannot be proved that the extensive Alaskan glaciation was contemporaneous with that of the Glacial Period in Europe and northeastern America, though it probably was is its inception. If this is so, then climatic change is without doubt the cause, for whatever the reason for the change, the spread of continental glaciers down into the temperate latitudes was a response to climatic change. Although we believe that the inception of the Alaskan glacier advance was contemporaneous with the spread of the continental glaciers in Europe and America, we are convinced that its termination was later than the period of recession of the continental glaciers. In a sense the Glacial Period is still in progress here, its influence being extended and prolonged by the lofty mountains, which have, in fact, partly or mainly reared themselves during the period of glaciation. Judging by the forest growth and by the extent of stream erosion since the recession of the ice from its most advanced position, we infer that the ice in its expanded stage lingered here until a few centuries ago.

One of our principal reasons for believing that the spread of glaciation in the Alaskan region dates back to the Glacial Period is that it has performed such vast work. Not only was a broad plain of glacial deposit built out into the Pacific beyond the mountain front, and an extensive submarine moraine built across the mouth of Yakutat Bay, but the topography was greatly altered by profound glacial erosion. This work must have required a long time. We cannot apply to this argument definite measurements from rate of ice erosion, but even the most extreme believer in the efficacy of ice erosion must agree that to lower a valley bottom from one to two thousand feet demands a great lapse of time. The main valleys have been greatly deepened; the lateral valleys have been deepened also, though to less extent; divides have been lowered and through valleys have been opened; the mountains have been sharpened; and, in fact, the topography has been completely altered in many of its most significant features by the glacial erosion resulting from the long flow of great ice streams. Thousands of years must be required for such vast work, and probably tens of thousands of years. It is by no means improbable that the beginning of extensive glaciation here dates back to the commencement of the Glacial Period, and that there was a long period of complex conditions of advance and recession with glacial and interglacial epochs. Of this, however, there is no proof. The period of glaciation was certainly many times longer than the period that has elapsed since the recession from the outermost stand.

When the glaciers were most advanced they literally flooded the mountain valleys,

and ice streams flowed across divides now discharging glaciers in two directions. Where fiord water now stands great glaciers 2000 or 3000 feet deep flowed through the fiord valleys (Pl. XCII, A); and where even the great glaciers, like the Hubbard, now terminate, the ice rose 1000 to 2000 feet higher than at present. Even such great ice streams as the Hubbard, Turner and Nunatak Glaciers are mere dwindling glaciers compared to their predecessors on the same sites. It was a region of mighty glaciers, flowing irresistibly along mountain valley courses, eating away the valley walls and scouring the valley bottoms; and this condition extended along the Alaskan coast from British Columbia to the Alaska Peninsula.

After its long, but unknown period of advance, the glacier system began a period of recession whose detailed history is also unknown. There may have been minor advances and recessions, or it may have been one continuous period of recession. By it the glacier fronts were caused to recede even farther than at present, and the forest extended farther up the fiord than now, and even into the region where glaciers now stand. From this we infer that both the length of time of recession and the extent of recession were greater than the present period of recession. But by this it is not meant to intimate that the period of recession was a really long one, for little gorge erosion has been accomplished by the streams issuing from hanging valleys that were produced by the ice erosion of the preceding great advance. A couple of centuries, or perhaps twice that amount, would suffice to explain the phenomena observed, and much more than that would be too great a time for the small work of stream erosion accomplished. Whether a similar period of recession was in progress in other portions of Alaska, or whether it was confined to the Yakutat Bay region and vicinity cannot now be stated.

During this period of recession, and perhaps mainly during the advance of the glaciers which terminated it, there was extensive deposition of gravels in Yakutat Bay, and especially in Russell Fiord. Then came a notable advance, pushing the glacier fronts once more far into Russell Fiord (Pl. XCII, B) and Disenchantment Bay, but not nearly out to the point reached by the earlier great advance. It was of brief duration, and the work performed was not notable. The valley walls were freshly scoured, and probably there was effective glacial erosion in parts of the fiord bottoms; but the ice erosion did not suffice to remove the gravels deposited between the two advances, though it greatly sculptured them. This advance was a local phenomenon, not occurring in Prince William Sound to the northwest, nor in the Inside Passage, though a similar, and probably contemporaneous advance is recorded in the Muir Glacier region and north of the St. Elias Range. The advance in the Yakutat Bay region reached its maximum in a very recent period, for the vegetation growing on a lake beach formed by the advanced glacier at the head of Russell Fiord is not over a half century old.

Following this notable, though brief advance came the recession which was still in progress at the time of our studies in 1905 and 1906. The recession was so rapid that vegetation had not been able to follow it closely, the overridden gravels near the glaciers having only scattered plants, while even the outermost portions of the overridden gravels bore only a young growth of cottonwood and alders. It is certain that the glaciers began recession from their outermost stand not over a century ago and perhaps not much more than half a century ago. The greatest recession has occurred in the tidal glaciers, notably the Nunatak, and in the Hidden Glacier. The least recession has occurred in the piedmont glaciers and piedmont bulbs. In fact, it is possible that the piedmont con-

dition is a direct result of the recent advance, lingering here because of the protection given by the sheets of ablation moraine by which the piedmont bulbs are covered.

The last episode in the glacial history of the region is the advance of the glaciers in response to the earthquake shaking in 1899, first observed in 1905, though affecting at least two of the glaciers before that time, causing the advance of four glaciers in 1906, one between 1906 and 1909, one in 1909 and one in 1910, and possibly slight advances of two others in previous years. Whether the advance has reached its maximum, or whether there are other and still greater changes in store, only the future can tell.

CHAPTER XII

THE GLACIERS AND GLACIATION OF PRINCE WILLIAM SOUND AND THE LOWER COPPER RIVER

Location and General View

The region of Prince William Sound and the lower Copper River is a little more than 200 miles west of Yakutat Bay and chiefly on the southern slope of the Chugach Mountains, an extension of the St. Elias Range. This part of the Chugach Mountains is from 50 to 75 miles wide and between six and eight thousand feet high, with a few loftier peaks, the highest of which, Florence Peak and Mt. Gilbert, reach elevations of 11,190 and 10,194 feet respectively. There is the same west northwest—east southeast trend as near Yakutat Bay and Mt. St. Elias, but, at the northwestern corner of Prince William Sound, the Chugach Mountains swing westward and come to an end. Thence the Kenai Mountains trend southward, forming the western side of Prince William Sound.

The Chugach Mountains are crossed by the canyon of Copper River, the greatest glacial stream in Alaska and one of the few rivers that completely cross the coast range. At its mouth it has built a great delta over 25 miles wide, a broad tract of low-lying coast, but west of the Copper River delta the mountains come down directly to the sea, and just here Prince William Sound penetrates the coast.

Prince William Sound (Pl. XCIII) extends about 80 miles east and west and about 50 miles north and south, and has numerous long bays and fiords extending even more deeply into the mountains. It is 80 or 90 miles from the head of ocean steamship navigation at the end of the Port Valdez fiord out to the coast where Hinchinbrook and Montague Islands guard the mouth of Prince William Sound from the open Pacific. It has a very irregular coast with long peninsulas reaching forward, between which bays and fiords extend back into the mountains. On the eastern side are Orca Inlet, Orca Bay, Port Gravina, Port Fidalgo, and Valdez Arm—Port Valdez. These are fiords from 12 to 30 miles long with mountainous peninsulas between. Hawkins and Hinchinbrook Islands correspond to these peninsulas but have been cut off by submergence or ice erosion. Besides these there are numerous small islands.

On the northern side of Prince William Sound are Columbia Bay, Long Bay, Wells Bay, Unakwik Inlet, Passage Canal, Port Wells and several smaller fiords. College Fiord and Harriman Fiord are tributaries of Port Wells. All these fiords are separated by long peninsulas and by islands, of which Glacier and Esther Islands are the largest.

On the western side of Prince William Sound south of Passage Canal are Port Nellie Juan and Icy Bay, separated by peninsulas and guarded by Knight, Latouche, Naked, and several other large islands and many smaller ones.

The lower Copper River is entirely in the mountains except for the broad delta at

its mouth. The topography of Prince William Sound is wholly mountainous. The greatest elevations are in the interior of the Chugach and Kenai Ranges and the peninsulas which project into the sound slope gradually to moderate heights and then end abruptly. There is an exceedingly small percentage of lowland. All of the islands are hilly, and the larger ones, like Knight and Latouche Islands, are mountainous.

The topography, geology, geography, and glaciation of Prince William Sound and the lower Copper River have been described by Schrader,[1] Mendenhall,[2] Schrader and Spencer,[3] Gilbert,[4] Emerson and Palache,[5] Gannett,[6] Burroughs,[7] Muir,[8] Keeler,[9] Fernow,[10] Davidson,[11] Grant and Higgins,[12] and others whose work is referred to in Chapters XIII to XXIII.

The geology of the Prince William Sound and lower Copper River regions may be summarized as follows. In Prince William Sound Grant [13] states that there are two groups of rocks. The Valdez group is older, perhaps Paleozoic, and consists of slates and graywackes. These are intensely folded in places and are somewhat metamorphosed. They contain granitic intrusions. The unconformable Orca group, probably Mesozoic, is made up of slates, graywackes, and rarer conglomerates and limestones. Locally there is much greenstone, made by the alteration of basic lava flows which were interstratified with the slates and graywackes. These are cut by bosses and dikes of granite and by basic dikes. Northern and western Prince William Sound, including all of the fiords except Ports Fidalgo and Gravina and Orca Inlet, are underlain by the rocks of the Valdez group while the islands and eastern shore of the sound, from Ellamar southward, are made up of the Orca group. It seems possible that the excavation of this great arm of the sea is related to its being floored by the relatively less resistant rocks of the Orca group.

In the lower Copper River, Hayes,[14] Schrader and Spencer,[15] Moffit and Maddren [16] have described the geology. There are three rock groups, (1) the Mesozoic Orca group

[1] Schrader, F. C., A Reconnaissance of a Part of Prince William Sound and the Copper River District, Alaska, in 1898, 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 347–423.

[2] Mendenhall, W. C., A Reconnaissance from Resurrection Bay to the Tanana River, Alaska, in 1898, 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 271–340.

[3] Schrader, F. C. and Spencer, A. C., The Geology and Mineral Resources of a Portion of the Copper River District, Alaska, House Doc. 546, 56th Congress, 2nd Session, Washington, 1900, pp. 9–92.

[4] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, Glaciers, 1904, pp. 71–97, 173-176.

[5] Emerson, B. K. and Palache, Charles, Harriman Alaska Expedition, Vol. II, Geology, 1904, pp. 24–26, 50–51.

[6] Gannett, Henry, Harriman Alaska Expedition, Vol. II, 1900, pp. 262–263; Nat. Geog. Mag., Vol. X, 1899, pp. 510–512; Bull. Amer. Geog. Soc., Vol. XXXI, 1899, pp. 346–348, 354–355.

[7] Burroughs, John, Harriman Alaska Expedition, Vol. I, 1901, pp. 63–76.

[8] Muir, John, *Ibid.*, Vol. I, 1901, pp. 125–127, 132–133.

[9] Keeler, Charles, *Ibid.*, Vol. II, 1901, pp. 219–221.

[10] Fernow, B. E., *Ibid.*, Vol. II, 1901, pp. 244, 248, 253.

[11] Davidson, George, The Glaciers of Alaska that are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc., Pacific, 2nd series, Vol. 3, 1904.

[12] Grant, U. S. and Higgins, D. F., Reconnaissance of the Geology and Mineral Resources of Prince William Sound, Alaska, Bull. 443, U. S. Geol. Survey, 1910; Tidewater Glaciers of Prince William Sound and Kenai Peninsula, Bull. U. S. Geol. Survey (in press).

[13] Op. cit. p. 11.

[14] Hayes, C. W., Nat. Geog. Mag., Vol. IV, 1892, pp. 126, 141–142, 154.

[15] Schrader and Spencer, op. cit.

[16] Moffit, F. H. and Maddren, A. G., Bull. 374, U. S. Geol. Survey, 1909, Pl. I.

south of Childs Glacier,[1] (2) the Paleozoic Valdez group throughout the Copper River canyon to a point between Tiekel River and Spirit Mountain, and (3) a belt of metamorphic slate, schist, and limestone, with intrusive diorite and greenstone, the limestone being in part of Carboniferous age.[2] The highest peaks and the deepest, narrowest portion of Copper River canyon are at the north in this third belt of most resistant rocks, and the broadest portion is at the south in the belt of weakest and youngest rocks.

Outside of Prince William Sound the Gulf of Alaska is relatively shallow, the hundred fathom curve being forty to seventy miles offshore. Off the Copper River delta it is easy to understand why one must go about 20 miles from shore before a depth of 350 to 400 feet is reached; but off Hinchinbrook Island, where there is no such active deposition, it is also necessary to go 5 to 10 miles before a depth of 300 to 350 feet is reached. Between the mouth of Prince William Sound and Middleton Island, which is 56 miles offshore, the depth of the Pacific Ocean averages only 200 to 450 feet.

In the straits at the entrance to Prince William Sound, that is between Hinchinbrook and Montague Islands, the water is 600 to 1300 feet deep and inside Prince William Sound the water averages 600 to 2400 feet in the main sound and 400 to 1400 feet in the tributary fiords, a striking contrast with the shallower ocean outside and one that will be explained later by the conditions of glaciation. One fiord, Orca Inlet, which opens to the Pacific, is only 6 to 60 feet deep and is quite unnavigable at the mouth, making it necessary for ocean steamships bound for the town of Cordova to go over fifty miles out of the way to reach the head of Orca Inlet by way of the main Prince William Sound and Orca Bay.

There are a few small native settlements on Prince William Sound and mining settlements on Latouche and Knight Islands and an old Russian trading post, Nuchek, on Hinchinbrook Island. At the head of Port Valdez, near Shoup and Valdez Glaciers, are an army post, Fort Liscum, and the town of Valdez, from which a military road extends into the interior of Alaska. Between the mouths of Valdez Arm and Port Fidalgo are Ellamar and several other mining settlements. There is a salmon cannery at Orca and nearby on Orca Inlet is the town of Cordova. The Copper River and Northwestern Railway extends eastward from Cordova, crosses a mountainous peninsula through the valley of Eyak Lake, and emerges on the Copper River delta, which coalesces with the deltas of the streams from Scott, Sheridan, and smaller glaciers in the mountains. The railway crosses Copper River just east of the old native settlement of Alaganik and goes up the eastern side of Copper River to Childs Glacier, which enters from the west nearly opposite Miles Glacier, which comes from the east. The railway crosses the river between these two glaciers and thence, after passing the Abercrombie Rapids, extends over the stagnant terminus of Allen Glacier, which enters Copper River valley from the west. Beyond this the railway follows the western side of the Copper River canyon, at the northern end of which is the town of Chitina.

Climate and Glaciation. Such precipitation records as we have in the Prince William Sound region show an annual rainfall of 190 inches at Nuchek at the western end of Hinchinbrook Island; at Orca of 149 inches; and at Fort Liscum an average of 76 inches. The precipitation is heaviest in fall and winter and, therefore, much of it is snow. In this latitude, 60° 30′ to 61° N., the warm waters of the Gulf of Alaska amelio-

[1] Brooks, A. H., Bull. 480, U. S. Geol. Survey, 1911, Pl. III and pp. 48–49.
[2] Moffit, F. H., Bull. 520, U. S. Geol. Survey, 1912, Pl. V and p. 94.

rate the climate, but they also supply moisture to the winds which blow upon the Chugach Mountains, giving rise to extensive precipitation there also, though of unknown amount, with a large proportion in the form of snow. The snow line is at about 3000 feet. There are, therefore, great snowfields covering all the upland areas where the slopes are not too steep. These snowfields feed glaciers which extend down the valleys to or near sea level. Most of the passes are filled with ice, but the glaciers are now far less extensive than formerly, as will be shown later, and indeed much less extensive than farther east where the snowfields of the loftier, eastern portion of the Chugach Mountains feed the piedmont Bering Glacier, and the snowfields of the still loftier St. Elias Range support the Malaspina Glacier and the glaciers of the Yakutat Bay region. The glaciers of the region to be discussed include various types from the smallest ice patches to great valley glaciers like the Miles and Columbia. Some of these are tidal, some end in the rivers, and some far up the mountain valleys.

Topography also controls the glacier forms in the Copper River valley where there are many excellent illustrations of the bulb type of glacier. Of these the Miles, Childs, Allen, Heney, and Sheridan are large and will receive more attention than the many other good-sized ice tongues.

In contrast with Yakutat Bay, which has only three glaciers discharging icebergs, there are 21 tidal glaciers in Prince William Sound. Their distribution is determined wholly by the topography and climate, for the largest glacier, the Columbia, is near the base of Florence Peak where the highest portion of the Chugach Mountains is close to sea level. Of the other tidal glaciers the largest number in a single arm of the sea is found in College Fiord where the Harvard, Yale, Smith, Bryn Mawr, Vassar, and Wellesley Glaciers reach tidewater. There are nearly as many in the loftier portion of the mountains near Mts. Gilbert, Gannett, and Muir, where the Barry, Harriman, Cataract, Surprise, and Serpentine Glaciers enter Harriman Fiord. Unakwik Inlet and Port Valdez in northern Prince William Sound have only one tidal glacier each, the Meares and Shoup, because the mountains are lower and the snowfall less. South of Port Valdez the eastern portion of Prince William Sound has no tidal glacier because the mountains are lower. South of Harriman Fiord the western portion of the sound has tidal glaciers only in a branch of Passage Canal, where the Blackstone and Ripon Glaciers extend into the water, in Port Nellie Juan, where the Ultramarine and Nellie Juan Glaciers reach the sea, and in Icy Bay, where the Chenega, Tiger's Tail, and Tiger Glaciers discharge icebergs. This is because the Kenai Mountains are lower toward the south.

Extent of Glacier Studies. The glacier observations in this region have resulted from various enterprises. The Russians who, like the Alaskan natives, attempted to ascend Copper River found the glaciers only a barrier, and, therefore, were not interested in them and have given us only a little information concerning the condition of these glaciers in early times. This, though natural, is unfortunate, but one cannot help regretting the absence of glacier observations in the time of the partial ascents of Copper River by Nagaief in 1783 and by other Russian explorers and traders in the Copper River and Prince William Sound regions. An account by Grewingk is the most valuable of the Russian descriptions of the Copper River glaciers. Some useful facts concerning Prince William Sound are supplied by Petroff. In this early period the charting of shorelines by explorers like Portlock in 1787, Fidalgo in 1790, and Vancouver, Whidbey and Johnstone in 1794, resulted in more definite information about the glaciers,

but only of the tidewater ice tongues. The same is true of the explorations of Applegate in 1887. Davidson has summarized many of these early explorations.

The explorations by United States army officers have supplied a little more information about the glaciers, because these parties were engaged in making maps. Therefore the expeditions by Abercrombie in 1884, by Allen in 1885, by Schwatka accompanied by the geologist Hayes in 1891, by Glenn, assisted by Castner, in 1898, by Abercrombie in 1898 and 1899 have furnished maps, locations, photographs, etc., of the Miles, Childs, Valdez, Yale, Portage, and other glaciers.

The utilization of the Valdez Glacier as a highway by thousands of prospectors in 1898, 1899, and 1900 added little to our knowledge of this glacier, although many men must have taken photographs and made observations which are not now available.

Then and in the years since, geologists like Hayes, Schrader, Rohn, Mendenhall, Spencer, Moffit, and others were visiting these regions, accompanied by army officers like Lowe and Babcock and the topographers Mahlo, Witherspoon, Gerdine, and others; and the maps, photographs, and observations by these men have been of great service in our glacier studies.

In 1899 the Harriman Expedition came to the Prince William Sound region to study the glaciers specifically and Gilbert's three days study and mapping of Columbia Glacier are of the utmost importance. He also made observations in College Fiord and Harriman Fiord, where Gannett made excellent maps and Merriam and others took glacier photographs of great value.

Two observers living within this area have studied the glaciers out of interest in their behavior, and Camicia of Valdez deserves great praise for his measurements of the retreat of Valdez Glacier between 1898 and 1911, as does Johnson of the Copper River railway for his studies of Childs and Miles Glaciers in 1908 and succeeding years.

The same is true of the studies of the Prince William Sound glaciers during parts of the summers of 1905, 1908, 1909, and 1910 by Grant, in connection with his U. S. Geological Survey work. During part of the time he was assisted by Higgins, some of whose glacier maps are reproduced in this volume. Grant and Higgins have written an account of the glaciers and glaciation of Prince William Sound. This had not been published at the time of completing this manuscript, although after doing our own field work in 1904, 1909, 1910, and 1911 we have seen the magazine articles containing abstracts of portions of their work. Grant generously allowed us to use his manuscript maps and unpublished photographs in the field, rendering our own field work much more effective.

The account here presented is based upon the earlier work, especially that published by Gilbert and the Harriman Expedition, by Grant and Higgins, and by Davidson, in addition to our own field work. This field work included (1) brief observations of Valdez Glacier by the junior author in June and September, 1904, (2) a few days' study by both authors of this book in August, 1909, (3) a little over two months detailed study and mapping by the junior author between June and September, 1910, and (4) a few days' observations in June, 1911.

Fifty-three of the glaciers of Prince William Sound will be discussed first, after which attention will be given to seventeen of the glaciers of the lower Copper River and to the phenomena of general glaciation in connection with these seventy larger ice tongues and certain smaller ones.

CHAPTER XIII

THE VALDEZ AND SHOUP GLACIERS

Location and General Relationships

Location. The glaciers discussed in this chapter are situated in northeastern Prince William Sound, upon the long narrow fiord called Port Valdez (Pl. XCIII), already alluded to as the northernmost of the fiords on the east side of Prince William Sound, where it penetrates most deeply into the Chugach Mountains.

The Port Valdez Fiord. The Port Valdez fiord extends northward from the northeast corner of Prince William Sound, between Glacier Island and Bligh Island just northwest of Ellamar, where it has a width of ten miles and a depth of 600 to 1300 feet. This lower portion of the fiord has Galena, Jack, and Sawmill Bays as tributaries, and narrows rapidly toward the northeast, until it attains a width of less than seven-eighths of a mile at Valdez (or Stanton) Narrows, where the depth is less than 500 feet and where a reef (Middle Rock) rises to the surface with surrounding depths of 180 to 300 feet. Here the fiord turns abruptly eastward, receiving Shoup Bay on the north side, widening to between three and four miles, and increasing in depth to 600 and in places over 800 feet. The southwest portion, outside the Narrows, is twenty miles long, the eastern part eleven or twelve miles. The shores have many minor capes, coves and small inlets and are mostly rocky.

At the eastern or extreme upper end Port Valdez shallows and there is a fringe of broad tide flats where Lowe River comes in from the east and Robe River and a number of other glacial streams from Corbin and Valdez Glaciers on the northeast. The narrower, unsubmerged valley of Lowe River extends the fiord valley eastward for twelve miles to the mouth of Keystone Canyon, beyond which the river heads against the Copper River headwaters on Thompson and Marshall passes. Other streams entering Port Valdez, beside those from the Valdez and Shoup Glacier valleys, are Mineral Creek, Fall or Gold Creek, and Canyon Creek, on the north, and Solomon Gulch and other unnamed creeks on the south. Several of these streams are in hanging valleys (h, Fig. 26).

The mountains enclosing the fiord rise four or five thousand feet within a mile or two of the shore, and five or six thousand feet a short distance further back. These mountains encircling Port Valdez maintain permanent snowfields which feed minor ice tongues. Of these there are three east of Solomon Gulch, Annin Glacier and three others between Valdez Narrows and Shoup Glacier, one in Fall or Gold Creek, the Corbin and Hogback Glaciers east of Valdez, and the Valdez and Shoup Glaciers which are to be more fully described.

The town of Valdez is at the extreme eastern or upper end of the fiord near the north side; and opposite it on the south shore, near the site of the former settlement called Swanport, is Fort Liscum, a United States army post.

The Valdez Glacier

General Description. The Valdez Glacier (Pls. XCIV, A, and XCV) is a through glacier whose northern terminus is called Klutena Glacier. It has been described by Schrader as follows.[1] "Starting from Valdez, the trail leads 4 miles northeast, with a very gentle rise over the delta gravels, to the foot of the Valdez glacier, thence about north for 18 miles up the glacier to the summit, which is 4800 feet high. The glacier is broken or transversely marked by four or five successive long benches or terraces, from one to the other of which the rise of 100 feet or more is usually sharp and sometimes difficult, the topography of the ice being rugged, with crevasses, ridges, and turrets. With the exception of these benches the ascent from the foot of the glacier to near the summit is gradual; but just before reaching the top there is a steep rise of a thousand feet at an angle of 15° to 20°. The pass is guarded by a couple of prominent peaks, one on either side and standing about a mile apart. From the summit the trail descends rapidly, but nowhere abruptly, for a distance of 6 miles through a canyon-like valley to the foot of the Klutena glacier, which is the source of the Klutena river."

Valdez Glacier trends a little west of north to the divide, Bates Pass, beyond which the other end of the through glacier bends eastward again. It averages two miles in width, while Klutena Glacier is from one to two miles wide. Omitting two barely detached tributary glaciers, Valdez Glacier receives two tributaries from the east and one from the west, while Klutena Glacier receives two eastern and one western tributary near Bates Pass. The latter, which has a width of about two miles, heads at least ten miles farther to the west on a 5000-foot pass, perhaps as a through glacier with the Shoup or Columbia ice tongues, and flows eastward and bifurcates near Summit Peaks, sending a smaller glacial distributary northeastward to Klutena Glacier and a larger one southeastward to Valdez Glacier.

The Bates Pass ice tongue is another through glacier east of this one (Pl. XCV), the divide being covered by a snowfield. The Klutena terminus is 2020 feet above sea level and from it a stream flows to Copper River. The Valdez terminus is 210 feet above sea level, the difference in elevation of the termini being mainly a response to the difference in snowfall on the seaward and landward sides of the mountain range. The mountain walls of the Valdez-Klutena through glacier rise from four to over seven thousand feet, being much oversteepened by former glacial erosion. The glacier surface is made up of a series of ice falls and gentler slopes so that the prospectors spoke of it as divided into "benches." There are no pronounced medial moraines, but the lateral moraines are broad, and near the terminus several faint medial moraines are disclosed by ablation.

The width of Valdez Glacier at the terminus is a mile and a quarter, or a mile and three-quarters around the slightly curved periphery of the front. The ice front rises 300 feet in the first quarter of a mile, and reaches an elevation of 1000 feet seven-eighths of a mile from the end, 1500 feet a mile and a half back, and 2750 feet at a distance of eleven and a half miles from the terminus.

A mile and a half from the end of the glacier, on the eastern side, is the valley of a detached tributary into which Valdez Glacier bulges slightly. The small glacier in this valley (which we have named Camicia Glacier) is three-eighths of a mile wide and

[1] Schrader, F. C., Report on Prince William Sound and the Copper River Region, in Maps and Descriptions of Routes of Exploration in Alaska in 1898, Special Publication, U. S. Geol. Survey, 1899, p. 61.

has one south tributary. Its end is heavily covered with moraine and completely disconnected from Valdez Glacier.

Between the terminus of Valdez Glacier and Port Valdez are three and a half miles of outwash gravel plain with many occupied and abandoned stream courses, a few knobs of dissected outwash gravels and recessional moraines, and, near Valdez, a strip of forest.

Observations Prior to 1898. Vancouver's lieutenant, Whidbey, entered Puerto de Valdes on June 10, 1794, making the first map of it and showing its head about half way between Valdez Narrows and the present terminus. He sailed northeast for twelve miles, or to what we now call Valdez Narrows, where he states[1] that a "small brook, supplied by the dissolving of the ice and snow on the mountains, flowed into the arm, which about five miles from thence terminated in an easterly direction." The latitude given and the shape of the fiord are essentially correct except that it does not extend eastward as far as now; but no glaciers are mentioned. Whidbey speaks of the region being in the same latitude as what we now call Columbia Glacier where they had observed great falls of ice, and of their surprise in Port Valdez that "in this branch no ice has been seen, notwithstanding it is terminated by shallow water at its head, and is surrounded by similar steep frozen mountains." Davidson has looked into Whidbey's description and believes[2] that if the shore on Whidbey's map "really was the limit of the bay, and the water was found shallow, then the whole eastern half of the Port, with its one hundred and fifty fathoms depth, was occupied by a low moraine-covered glacier that hid the ice front."

We are not convinced of the correctness of this interpretation, partly because no thundering ice falls or floating icebergs are mentioned, as would be necessary with the existing depth of the fiord, partly because the original account does not prove that Whidbey went far enough through the Narrows to see the whole length of Valdez Arm and the Shoup and Valdez Glaciers, but chiefly because eastern Prince William Sound does not seem to have had such a recent episode of ice advance as is shown in Yakutat and Glacier Bays. An eight or nine-mile retreat for Valdez Glacier between 1794 and 1898 is not impossible, but, if Davidson's interpretation is correct, the vegetation has followed the ice front much more closely here than in similar cases we have observed in Alaska, for in 1909 spruce was growing at Valdez and cottonwood and alder extended right up to the glacier. Moreover, if Valdez Glacier was so extensive in 1794, even if moraine-covered and not discharging icebergs, and not recognized as a glacier, Shoup Glacier should have been similarly expanded and must have been visible to Whidbey from the Narrows, should have discharged icebergs, and should have spread over slopes that are now too well tree-covered to have been reforested in the years between 1794 and 1898. Another objection to this theory of recent great expansion of Valdez Glacier is the presence of the very extensive outwash gravel plain which extends from the Valdez Glacier front three and a half miles to the head of the fiord. A long period of time has been required to build this gravel plain, even assuming former shallow water where it stands; and, moreover, mature cottonwood trees have had time to grow upon it since it was built above sea level.

[1] Vancouver, Capt. George, A Voyage of Discovery to the North Pacific Ocean and Around the World, Vol. V, London, 1801, pp. 317–8.

[2] Davidson, George, The Glaciers of Alaska that are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc., Pacific, Vol. III, 1904, pp. 35–6.

We are inclined to think, therefore, that although Valdez Glacier is now receding from a fairly recent, more advanced position, the evidence is not conclusive that the advance was so recent as suggested by Davidson's interpretation of Whidbey's observations in 1794.

There is a rumor that the Russians went into the Copper River valley from the head of Port Valdez, but it is not known when, nor whether they went over the Valdez Glacier, nor what conditions they encountered.

Petroff stated [1] in a report written in 1882, that "in Port Valdez, at the northern extremity of the sound, a glacier exists with a face 15 miles in length at the sea shore, while its downward track can be traced almost to the summit of the Alps. Huge icebergs drop off its face with a thundering noise almost continually and drift out to sea, and the whole extensive bay is covered with small fragments, making it inaccessible to even boat navigation, and consequently a safe retreat for seals, which sport here in thousands. Port Fidalgo in the east and Port Wells in the west also have tremendous glaciers."

The presence of a glacier "fifteen miles in length at the sea shore" is manifestly impossible in Port Valdez fiord which could not have held such a glacier, even if Valdez and Shoup Glaciers had then coalesced; nor could such a large glacier have existed anywhere in Prince William Sound, for the expansion and union of Valdez, Shoup, and Columbia Glaciers would result in a united ice front of only ten miles. Vegetation and human records prove that no such expansion could have occurred between 1794 and sometime between 1868 and 1882, the dates between which any possible Russian observation to which Petroff might have had access must have been made. The dimensions are evidently an error, possibly of printing or of translation, for Petroff knew the Russian records as well as any man.

He specifically mentions the Port Wells glaciers to the west, which are still tidal, and glaciers in Port Fidalgo on the eastern side of Prince William Sound, where no glaciers are now known, but does not mention the great Columbia Glacier, the largest ice tongue of the whole region. We think it probable that the Columbia Glacier reached tidewater in the period mentioned, as it does now, and are, therefore, inclined to believe that the Columbia Glacier is referred to in the sentences quoted from Petroff.

In September, 1884, Lieut. W. R. Abercrombie of the U. S. Army entered Port Valdez and ascended one of the glaciers to a point where he could see a lake beyond.[2] It is not evident whether this was the Corbin Glacier or some other ice tongue that was then more extensive than now. It can hardly have been the Valdez Glacier and Klutena Lake, for the direction is wrong and the distance too great. The ice pass sought was one formerly (1850–60) used by the Copper River natives on their way to Prince William Sound. It may have been one leading to Woodworth Glacier in the Tasnuna Valley, though the distance is rather great and there is now no lake. It was not Marshall Pass, which has no glacier. Allen's map made in 1885, perhaps from Abercrombie's data, shows this route with the glacier and lake,[3] but not on a large enough scale for

[1] Petroff, Ivan, Population, Industries and Resources of Alaska, Tenth Census of the United States, 1880, Vol. VIII, 1884, p. 27.

[2] Abercrombie, W. R., Supplementary Expedition into the Copper River Valley, Alaska, In Narratives of Explorations in Alaska, Washington, 1900, pp. 391–2; Amer. Geol., Vol. XXIV, 1899, pp. 349–354.

[3] Allen, H. T., An Expedition to the Copper, Tanana, and Koyukuk Rivers, Senate Ex. Doc. 125, 49th Congress, 2nd Session, Washington, 1887, Map 2.

A. Shallow Crevasses of Hidden Glacier in 1909

B. Details of Surface of Hidden Glacier in 1909

A. HIDDEN GLACIER DELTA IN 1899 FROM PHOTO STATION C
View by G. K. Gilbert. The Harriman Expedition ship *George W. Elder* at anchor at low tide.

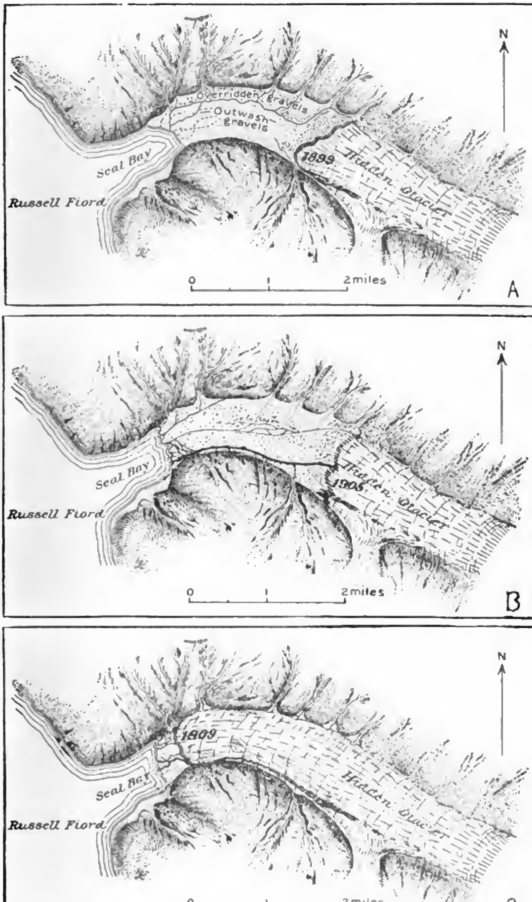
B. THE FRONT OF HIDDEN GLACIER DELTA IN 1910
Picture taken at low tide. Between 1899 and 1910 the front of this delta was built forward 1600 feet, so that the place where the Harriman Expedition ship was able to anchor at low tide in 1899 was dry land at all stages of tide in 1910. Both pictures from same site.

Plate LXXXIII

Northern Margin of Hidden Glacier in 1909

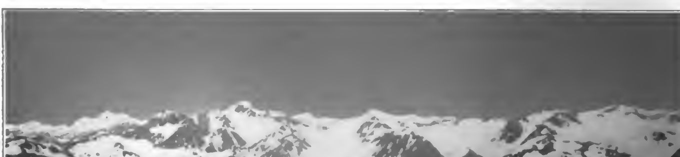
PLATE LXXXIV

MARGINAL STREAM ON NORTH SIDE OF HIDDEN GLACIER IN 1909

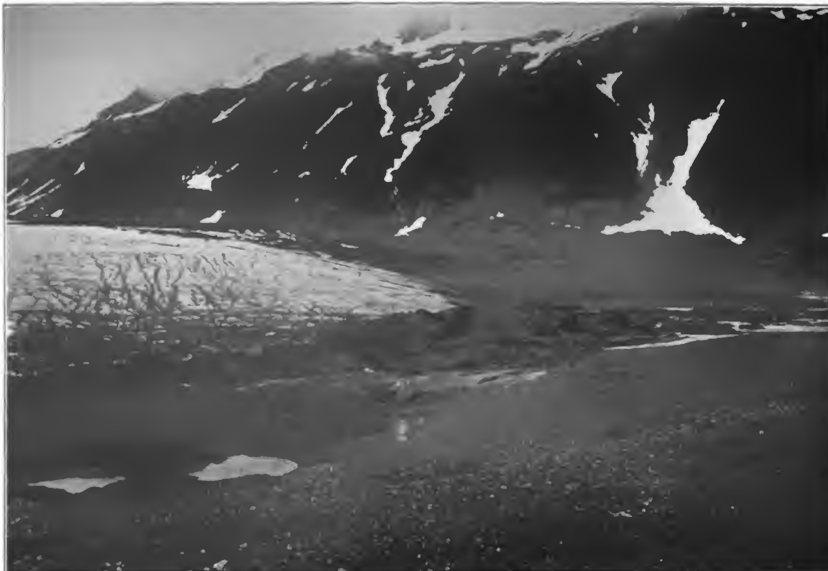


Hidden Glacier in 1899, in 1905 and 1906, and in 1909 (Topography after Gannett)

A. Northern Margin of Hidden Glacier in 1909
Vegetation cover on left. Barren area in middle and ice on right.



B. Fourth Glacier in 1906
Photograph by Canadian Boundary Survey.



End of Fourth Glacier in 1909
Smooth surface, débris-covered end, barren gravels, and deglaciated mountain slopes. Kettles in gravel area.

Plate LXXXVIII

End of Fourth Glacier in 1909
Photographed from 470-foot hill (Fig. 16).

A. ESKER ON OVERRIDDEN GRAVELS SOUTH OF CAPE ENCHANTMENT
On west side of south arm of Russell Fiord. Note general absence of vegetation. Photograph taken July 23, 1905.

B. MCCARTY GLACIER IN 1905

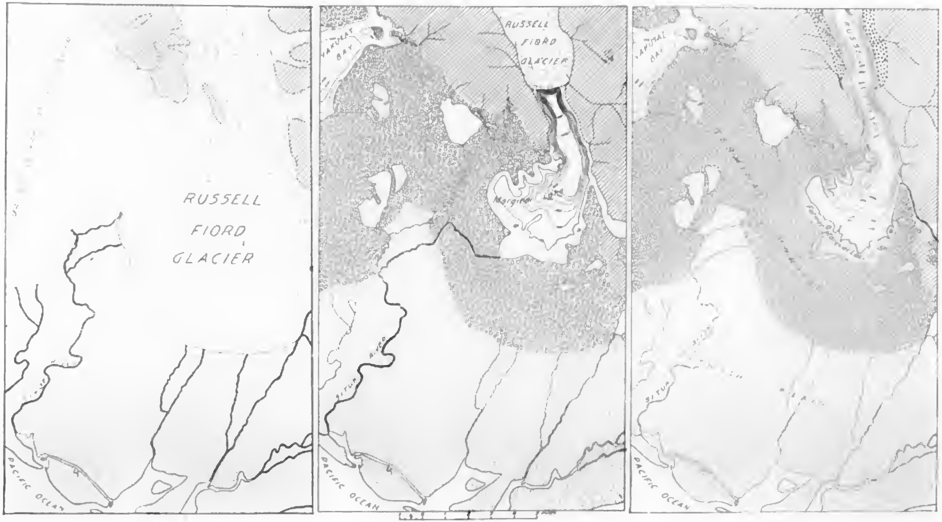
Plate XC

Ice Jam in Yakutat Bay in June, 1910

Plate XCI

Soundings and Submerged Contours in Russell Fiord and Disenchantment Bay

Plate XCII



A. B. C.

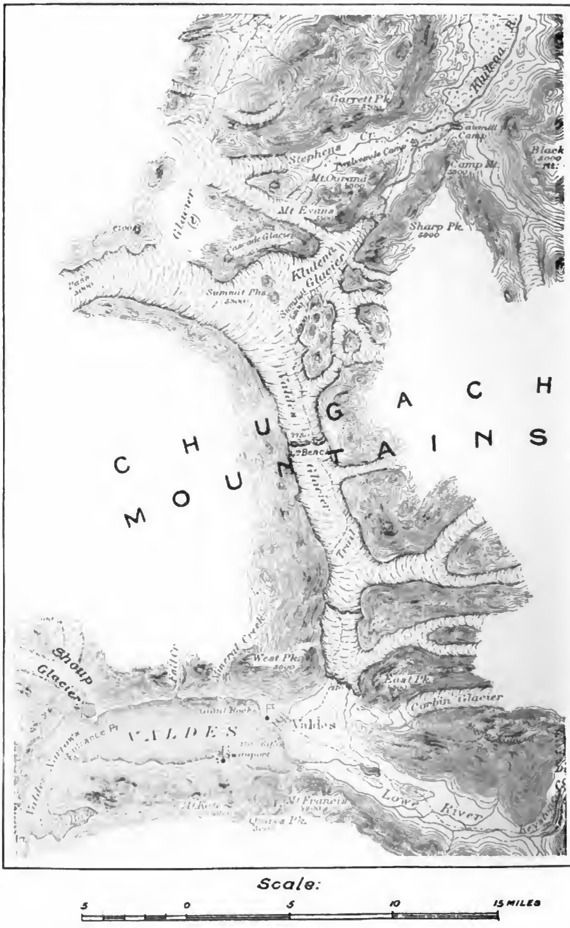
Three Maps Showing South End of Russell Fiord

At (a) maximum advance, (b) last recent advance, (c) at present, with existing areas of outwash, terminal moraine, former lake bottom, overridden gravels, and submerged moraine.

Plate XCIV

A. Valdez Glacier from the East Side
Photograph by A. C. Spencer, 1900.

B. The Clear Ice Surface of the Medial Portion of Valdez Glacier
View looking up the glacier. Photograph taken August 18, 1909.



THE VALDEZ GLACIER HIGHWAY (AFTER SCHRADER, MAHLO, AND LOWE)
An ice-filled pass across the Chugach Mountains, crossed by thousands of prospectors. The Shoup Glacier, used as an ice-house by the soldiers at Fort Liscum, near Swanport.

ICEBERGS ON SITE OF MARGINAL LAKE
Between Caunicia Glacier (in background) and Valdez Glacier (in foreground), August 18, 1909.

A. DÉBRIS CONE ON THE FRONTAL MARGIN OF VALDEZ GLACIER
Photograph, August 18, 1909.

B. WESTERN MARGIN OF COLUMBIA GLACIER (ON LEFT). AUGUST 23, 1909

identification of the glacier mentioned by Abercrombie. We are, therefore, unable to state specifically what the glacial conditions in Port Valdez were even as late as 1884.

Valdez Glacier between 1898 and 1911. In 1898 Valdez Glacier became a highway of travel into the interior of Alaska,[1] and from that date to the present we have definite knowledge about the general condition of the glacier. It is not known who was the first of the prospectors who made his way across the ice-filled pass of Valdez Glacier, on his way to the Klondike gold fields by the All-American route or to the Copper River valley, nor when the first traverse was made. Doubtless this first journey was made before 1898, for during February, March and April of that year three thousand people landed at Valdez, and one or two thousand more came during the summer. Some prospectors had reached the summit of Valdez Glacier by April 15th, and before May 3rd 2000 men had crossed the pass and gone down the Klutena Glacier, while 1500 more were on their way.[2] Mr. Charles Simonstad of Valdez who crossed the glacier in 1898 states that 5000 men landed that year, that 4500 crossed the glacier pass, and that all but two or three hundred of them returned that fall by the same route.

In April and May, 1898, some of the army detachments crossed the glacier pass several times. These men and the many prospectors encountered great hardships, being unprepared for glacier travel; but the generally inactive condition of the glacier may be inferred from the fact that three or four thousand men were able to travel over it as a highway during 1898, to transport thousands of pounds of provisions and outfit, and even to take pack animals over the twenty-four miles of ice, twenty-three horses and mules being used by a single one of the army parties, and fourteen by another. The geologist Schrader and several of the army officers have described the glacier with its ice falls, benches, and crevasses, the descriptions and photographs by Schrader and the map by Mahlo showing conclusively that the surface and the position of the terminus have fluctuated only slightly between 1898 and 1909. The 1898 map shows the glacier terminus projecting farthest near the eastern side of the valley and two streams flowing to the fiord, one on the western side through Valdez, the other by way of Robe Lake. This map is reproduced as Pl. XCV. A large photograph of the front of the glacier taken by J. H. Steiner on July 14, 1898, shows the western margin ending close to a distinctive talus cone on the mountain side, and exactly at, or within a few score feet, of the position of the same margin in 1909.

The second Copper River Exploring Expedition of the U. S. Army, in 1899, also under charge of Captain Abercrombie, gave little attention to the Valdez Glacier and very few prospectors crossed the glacier pass that year, although many had come back that way during the previous fall and winter. It was crossed by at least one army party,[3] however, J. F. Rice going over the glacier and back again; and the front was mapped by Lieut. W. C. Babcock, who also took many photographs of the ice front,

[1] Martin, Lawrence, Mastering the Alaskan Glacier Barriers, Scientific American Supplement, Vol. LXXI, 1911, pp. 305–307; Alaskan Glaciers in Relation to Life, Bull. Amer. Geog. Soc., Vol. 45, 1913, pp. 801–818.

[2] Abercrombie, W. R., Reports of Explorations in the Territory of Alaska, 1898, War Dept., Adj.-Gen. Office, No. XXV, Washington, 1899, pp. 299–300, 304–309, 343–344, 353–354, 389–391, 394–401, 403–407, 421–423, 433–435, 451–452, 454–455; also republished in Narratives of Explorations in Alaska, Washington, 1900.

Schrader, F. C., A Reconnaissance of a Part of Prince William Sound and the Copper River District, Alaska, in 1898, 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 350–353, 355–356, 381–382.

[3] Rice, John F., In W. R. Abercrombie's Copper River Exploring Expedition, Alaska, 1899, Washington, 1900, pp. 55–57.

published in the report referred to above, several of which may be used for determining the slight changes in the glacier terminus between 1899 and 1909. His map shows that the end changed from a southwest-facing terminus projecting most on the eastern side in 1898 to a south-facing terminus projecting most in the middle in 1899. The glacial streams changed slightly also.

Schrader and Spencer, in 1900, went into the Copper River valley by way of Valdez, and their topographers, Gerdine and Witherspoon, represented the Valdez Glacier upon a map, although their route lay over a new military trail which did not traverse the Valdez Glacier. They state that most of the prospectors who went into the interior

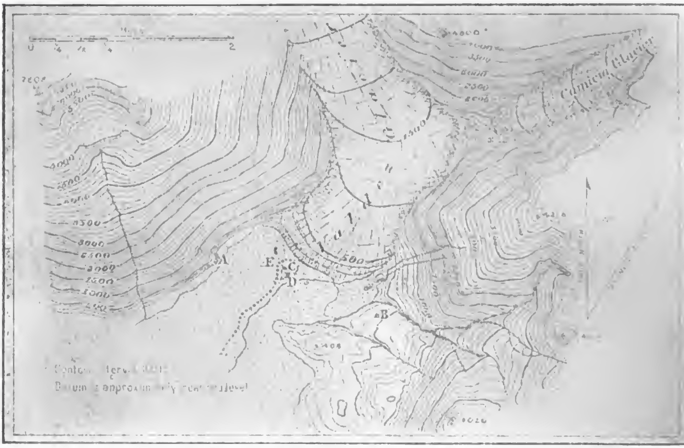


Fig. 22. Sketch Map of Valdez Glacier in 1909.

The contours are in error, owing to an accident to the instrument used.

over the snow early in 1900 followed the glacier route, but that it was not utilized during the summer.[1] Their map and a photograph by Spencer show the ice front much as in 1899, though the streams on the outwash plain were slightly different.

In 1901 begins the series of valuable observations of Valdez Glacier by the late L. S. Camicia, an optician and watch-maker at Valdez, who took pains to go to the glacier nearly every year since then, and who made careful measurements of the retreating ice front. He visited the glacier in 1898 and 1899 but made no measurements. In 1901 he selected stations and built monuments upon morainic knolls in front of the glacier, and, by pacing and measuring with a twenty-foot rope, determined

[1] Schrader, F. C. and Spencer, A. C., The Geology and Mineral Resources of a Portion of the Copper River District, Alaska, House Doc. 546, 56th Congress, 2nd Session, Washington, 1901, p. 19 and Plate II; also republished in Pl. I, Bull. 374, U. S. Geol. Survey, 1909, and Chitina Quadrangle, Map 601, U. S. Geol. Survey.

the distance from the stations to the ice front. These observations were repeated in the years listed below, showing that the glacier retreated 586 feet between June 6, 1901, and June 18, 1911, or an average of about 58½ feet per year. It is unfortunate that Dr. Camicia made no observations in 1906 or 1907 when the otherwise steady retreat was interrupted, as will be shown later. He states that the glacier has retreated more on the east than on the west side. The measurements given below are about in the middle of the ice front.

TABLE SHOWING RETREAT OF VALDEZ GLACIER DURING TEN YEARS

Date of Observation	*Amount of Retreat*	*Total Retreat*
June 6, 1901, to June 18, 1902	39 feet	39 feet
June 18, 1902, to June 7, 1904	165 "	204 "
June 7, 1904, to June 16, 1905	97 "	301 "
June 16, 1905, to Oct. 13, 1908	205 "	506 "
Oct. 13, 1908, to June 21, 1909	44 "	550 "
June 21, 1909, to June 18, 1911	36 "	586 "

In June and September, 1904, the junior author spent several days at Valdez and made brief observations of the glacier. The following description is based upon notes made at the time. The front had no terminal cliff but ended in a low sloping edge, almost as in 1909. The lateral moraines were much as they are now, and in a wedge near the center the ice was mantled with angular fragments for a half mile from the front, but was less crevassed than in 1909. The area of crevassing near the turn in the glacier on the west side was much as during the whole period of observations. There was a good-sized lakelet in front of the glacier, west of the middle, a very large stream flowing from the eastern margin of the glacier, and a smaller stream from the western margin. The latter was very much larger than the stream occupying the same position in 1909.

Valdez Glacier has been twice visited by Professor U. S. Grant. He made observations and took photographs in 1905 which enabled him to detect an advance between that year and 1908, which he describes [1] as follows:

"The western part of the front of this (Valdez) glacier was visited about August 1, 1905, and again on July 11, 1908. Some time during this interval the glacier has advanced 250 to 350 feet and built a moraine and then retreated nearly to its former position. On the extreme western edge the ice in 1908 was about 100 feet in advance of its position in 1905."

Since Dr. Camicia did not visit the Valdez glacier in 1906 or 1907, we cannot be sure in which of these years the advance took place. It must have been slight and brief, as it escaped observation by the people of Valdez, only three and a half miles away. A photograph taken in June, 1907, proves that the advance was not going on at that time, and it seems probable that it occurred in 1906.

[1] In H. F. Reid's Variations of Glaciers, Journ. Geol., Vol. XVII, 1909, p. 670.
Grant, U. S. and Higgins, D. F., Glaciers of the Northern Part of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLII, 1910, pp. 722–726.

Observations by the National Geographic Society. The National Geographic Society's expedition spent August 17–19, 1909, in studying and photographing Valdez Glacier.[1] We made slight additional observations in 1910 and 1911. In 1909 Mr. Lewis made the topographic sketch map reproduced as Fig. 22 and established the bench marks from which transit readings on the ice front were made. These bench marks, for Valdez Glacier, are located as follows.

Transit Readings. Transit readings from Stations A and B, Valdez Glacier, August 19, 1909.

Station A (Fig. 23) is located on the hillside on the west side of the alluvial outwash plain, one-half mile south of the glacier front. It is a stake driven two feet in the ground, just above a bare brownish cliff about 200 feet above the plain.

READINGS

Point Sighted	*Ver. A*	*Ver. B*	*Angle from A–B*
1. Station B	0° 00′	180° 00′	
2. Extreme southern point of glacier	349° 38′	169° 38′	10° 22′
3. Ice front about middle of glacier	346° 51′	166° 51′	13° 09′
4. West margin of glacier where ice disappears from view around hill. Point is about one mile north of ice front	301° 18′	121° 18′	58° 42′

Station B is located on the point of a spur east of the outwash gravel plain and about 1200 feet east of the present discharging stream of the glacier. A creek from the high hills to the east, running west, bears north of the station about 600 feet, and a prospector's cabin bears north about 1500 feet. The station is marked by a stake driven in a crack in a bowlder 4 by 6 feet on top.

READINGS

Point Sighted	*Ver. A*	*Ver. B*	*Angle from A–B*
1. Station A	180° 00′	00° 00′	
2. Ice front about middle of glacier (Same as No. 3, Station A)	195° 50′	15° 50′	15° 50′
3. Point on ice front half way between middle and west edge	197° 04′	17° 04′	17° 04′

Camicia Glacier. The first tributary valley on the east side of Valdez Glacier is occu-

[1] Nat. Geog. Mag., Vol. XX, 1910, pp. 7, 9, 11, and 22.

pied by the Camicia Glacier [1] which is detached from the main ice tongue, as already stated. Between this former tributary and the Valdez Glacier is a triangular depression into which the Valdez Glacier bulges slightly (Fig. 22) and this depression, one or two hundred feet deep, sometimes contains a marginal lake, like the Merjelen See on the border of the Aletsch Glacier in Switzerland. At the time of our visit in August, 1909, there was no lake present, but the numerous stranded icebergs attested its recent presence, the fact that these icebergs had not melted and that some of them rested upon last winter's snowbanks showing that the lake had existed in the spring of 1909. The lake

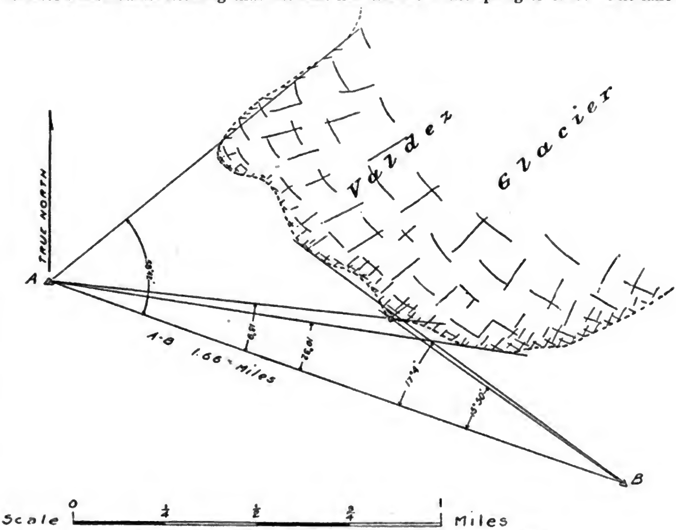


Fig. 23. Triangulation on Front of Valdez Glacier in 1909.
For location of stations A and B see Fig. 22.

occupied a triangular area of nearly half a square mile and probably received part of its water from a stream issuing from Camicia Glacier and part from the marginal drainage of Valdez Glacier itself. The lake possibly forms every spring or winter, when glacial advance and freezing closes crevasses and sub-glacial channels, and drains again early in the summer through a submarginal channel. That the draining of the lake of 1909 had been completed in August is shown by the icebergs (Pl. XCVI) in the picture reproduced here.

This detached tributary, which is three-eighths of a mile wide and has one south

[1] Named for the late L. S. Camicia of Valdez, whose valuable observations on Valdez Glacier have already been mentioned.

tributary cascading out of a lofty hanging valley, ends in a moraine-covered nose, beneath which a stream emerges from an ice cave. About two miles from the edge of Valdez Glacier the clear ice begins, and thence eastward to the summit the surface is bare of moraine. There are no medial moraines but there are broad lateral moraines, that on the south side being widest, partly because of the supply from the cascading tributary.

Surface of Valdez Glacier. The surface of Valdez Glacier is generally clear and free from moraine except along the margins and near the terminus. It is moderately crevassed (Pl. XCIV, B), but melting has rounded the seracs and reduced their height so that the crevasses are rather broad, and none of them deep. One can make fairly easy and rapid progress over the glacier surface by winding back and forth between crevasses. The surface is made up of coarsely-crystalline ice so melted along the crystal faces for the upper six inches that it is granular and crumbles as one walks over it. Pieces one or two inches square can be picked out and through each numerous holes and cracks extend.

The lateral moraines are a mere veneer of slabby, angular débris upon the surface. In the east lateral moraine, where the ice is somewhat crevassed, the veneer is so thin that sheer cliffs stand out as a result of melting, and the crevasses are only partly filled. Upon the lateral moraines and the débris-laden terminus are numerous lines of flat stones whose linear arrangement suggests that they represent the bottom of former crevasses. Some of the lines of these crevasse deposits now stand in relief. There are also numerous cones near the terminus of the glacier that appear to be piles of dirt but are really cones of ice (Pl. XCVII, A), very thinly veneered with fine black slate fragments, which have protected the ice and kept it from melting. Some of these are six or eight feet high, and some are low and in form resemble ant hills. Many of these smaller cones are also in lines and suggest moulin deposits along the bottom of crevasses. There are occasional glacier tables where flat rocks stand upon a column of ice which has been protected from melting. The material upon the glacier surface is predominantly slate and graywacke of the Valdez series with a few crystalline rocks, such as might come from the granitic diorite dikes of the area drained by the glacier and its tributaries.

Front of Valdez Glacier. The terminus of the glacier slopes gently (Pl. XCVIII, A) and is continuously dirty. Near the eastern side two rock knobs, each over a hundred feet high, rise at the edge of the glacier, being the only rock hills exposed away from the valley walls. East of these the largest stream emerges and the lateral moraine is wide, while to the west a V-shaped area of débris-covered ice extends far up the glacier. The west lateral moraine is also broad but is interrupted by the area of crevassing where the glacier turns sharply at the bend in the valley less than a mile from the end. The débris-covered portion of the terminus is narrowest in the middle of the glacier between the west lateral moraine and the medial moraine.

The streams emerging from the middle of the glacier front in 1909 were of small size and there were no good-sized frontal lakes, though there were several small ones. The stream on the west side was much smaller than that on the east. The drainage evidently shifts back and forth across the ice front, as is shown by the abandoned stream channels and terraces and the absence of vegetation upon the outwash gravels in front of the glacier. The largest stream has, however, remained on the east side of the glacier since 1898; there has sometimes been a stream in the center and sometimes not; and the stream on

the west side has fluctuated notably in volume. Part of this variation is doubtless due to fluctuation according to the season of the year.

Recession of the Glacier. Photographs taken in 1909 from sites occupied by Grant in 1905 and 1908 show that the retreat of the glacier is continuing, as is also shown by Dr. Camicia's measurements. Recent retreat is also attested by the presence of detached ice masses, partly covered with ablation moraine or with outwash gravels, some distance out from the ice front on the west side, and at several points near the center.

There is no continuous recessional moraine in front of the Valdez Glacier, the 468-foot ridge projecting from the east side of the valley, three quarters of a mile south of the glacier (Fig. 22) being a rock spur. West of the center of the glacier there is, however, an elongated morainic hillock 75 to 100 feet high. It is 850 feet from the ice front and extends parallel to it, the site of Dr. Camicia's original monument and of one of our camera stations (C, Fig. 22). On either side of it are two or three other hillocks forming a crescent. This monument ridge is made up of rounded outwash gravels and of angular scratched bowlders including many crystalline rocks, some of them large, and seems to contain no ice. The knolls to the west are chiefly dissected, outwash gravel remnants, while some of those to the east still contain ice. On these knolls are scattered shrubs ten years old or so, indicating that the year 1898, when the prospectors report some of these knolls uncovered, a condition also shown in photographs, was soon after the ice had left the monument ridge. Probably it even later abandoned some of the hills to the east which still contain ice. The rock knobs on the east side of the glacier have only annual plants growing upon them and have apparently been uncovered later still. The mountain slopes on either side of the glacier also have a barren strip at their base from which the ice has recently receded.

Outwash Gravel Plain. From the glacier to the fiord, a distance of three and a half miles, extends a broad, perfectly-developed outwash gravel plain, which descends sixty feet or so to the mile. It is made up predominantly of rounded outwash gravels deposited by the glacial streams, and has a high percentage of slate fragments. There are practically no large bowlders. Over the whole surface of the plain there are numerous dry channels and low terrace strips between channels. This outwash plain has an area of about ten square miles, extending southward to Lowe River. The active aggradation by the large stream from the eastern margin of Valdez Glacier, locally known as Glacier River, has built up the gravel plain so high that a lake is held in between the outwash gravel plain, the mountain side and an east-west ridge on the north side of Lowe River. This is Robe Lake, which is over two miles long and has had one of the glacial streams flowing into it since before 1898. The outlet, Robe River, receives as tributaries several other glacial streams (Pl. XCV).

The outwash gravels are slumping for a half mile from the ice front near the western valley wall, and there are kettles partly filled with water, showing that buried ice extends at least this distance out from the visible terminus. East of the center other areas of slumping show buried ice beneath the gravels, though these do not extend so much as a quarter mile from the terminus. The persistent rumor in Alaska that ice underlies the whole gravel plain, and even the site of the town of Valdez, and that it has been reached by wells and pulled up by the anchors of ships, is wholly without basis; at least we have not been able to find any definite evidence of its truth.

Near the glacier there is no vegetation upon the outwash gravel plain, except upon the

isolated knolls already mentioned and upon one strip of elevated outwash terrace which has been immune from stream encroachment since before 1898, and upon which shrubs grew at that time, as photographs show. East of this terrace there is no part of the outwash plain, except the isolated gravel knobs, that has any shrubs whatever. That streams flow over every part of this surface sometime during practically every year, is indicated by the fact that not even annual plants can secure a permanent foothold. One small strip near the center of the plain that had scattered shrubs in 1898 has since had them destroyed by the streams.

The above conditions apply to the mile or so of outwash plain nearest the glacier, where there are a very few small areas of scattered shrubs, but the next mile and a half or two miles, the width varying in different parts of the plain, has absolutely no vegetation except a few annual plants. The mile, or thereabouts, of plain nearest the fiord, on the other hand, has a predominance of vegetation cover, chiefly cottonwoods and alder, with narrow stream-channel strips extending through the forest. There are trees several feet in diameter, indicating a long period of immunity from extensive and continuous stream encroachment in this part of the outwash plain. This period of immunity is apparently now at an end, for there are lanes of stream channel extending through the forest, in some of which the trees have been removed while in others the dead trees, 40 to 50 feet high, still stand upright. The northern margin of this forest is made up of long points of tree growth, extending up the plain toward the glacier. That this finger-like inner margin is a condition due to the destruction of forest rather than to advance of trees is suggested by the recent killing of mature cottonwoods which had grown to a diameter of two or three feet far up on one of the points. That all the forest has not been removed by the streams may be due to lack of time, the streams possibly now being more active again after a period when the glacier was farther back than now. Or it may be due to the fact that the streams have become entrenched in the gravels at other points as far as this down the outwash plain and while they persist there, vegetation thrives elsewhere. Sometimes the streams do leave these channels and mow down a section of forest, as during the floods in 1905 when trees were uprooted or smothered, and even houses washed away in the town of Valdez. The forest advances farthest toward the glacier in the eastern side of the outwash plain where the largest stream has flowed ever since 1898, and where now many of the stream channels extend down through the forest. We are at a loss to account for this peculiar distribution of tree growth on the Valdez Glacier outwash gravel plain, though it is possible that the point of outflow of the largest glacial stream varies from time to time, the latest point, since 1898, being on the eastern side and the stream from it still being engaged in raising the depression of Robe Lake to the level of the previously-deposited central part of the plain.

Former Advance. The recency of retreat of Valdez Glacier from the latest notable advance is proved by the distribution of vegetation, the clearness of grooving on the valley walls, and the streams cascading down the mountain sides as yet little entrenched in gorges. The possibility of retreat of the glacier from this former stand to a point even farther than at present is suggested by the vegetation-covered talus slopes on the west side of the valley, of which the ones nearest the glacier began to be undercut by glacier and streams before 1898, so that part of the vegetation has been removed.

At the maximum of glaciation the main ice current of Valdez Glacier seems to have been forced over to the west wall by the thrust of the Corbin and two other tributaries from

the valleys to the east. The west wall is consequently oversteepened by glacial erosion up to 1500 feet or more, while on the east side a long rock spur three quarters of a mile south of the glacier was not completely removed, nor the minor rock knobs wholly erased. Even on the eastern side, however, glacial erosion has oversteepened the valley notably, leaving three valleys hanging above the main valley. Corbin Glacier and the one north of it end within their hanging valleys, while the northernmost glacier still cascades out over the lip of its hanging valley. It is not known what the relation is between the Valdez Glacier and Lowe River valley, for their bedrock bottoms are completely buried beneath the outwash gravels; but it is suspected that Valdez Glacier was the master ice tongue and that Lowe River valley, which at present has no glacier, hangs above it.

The Shoup Glacier

General Description. The Shoup Glacier,[1] at one time called Canyon Creek Glacier, enters Port Valdez just north of Valdez Narrows, terminating a little over two miles back from the main fiord. It has been erroneously represented upon several maps [2] as made up of two glaciers which coalesce two miles from the terminus. As a matter of fact there is no west tributary, but the Shoup Glacier is of a peculiar S-shape (Fig. 24). It comes from an unknown source in the Chugach Mountains, perhaps heading on a through glacier pass with the west arm of Valdez Glacier, in a region characterized by Schrader as " a waste of glaciers and névé." It flows southwest for an unknown distance and then bends sharply southeastward around a right-angled elbow (Pl. XCVIII, B) where, 1000 feet above sea level, it is a little over a half mile wide. Below this point it expands to a mile and a quarter and flows to the sea, where its terminus is three quarters of a mile wide. The mountains rise to heights of from 4000 to 6000 feet near the glacier, and the valley walls ascend precipitously from 1000 to 3000 feet. There is an ice cascade at the elbow, below which the glacier slopes smoothly nearly to the water's edge, and there descends steeply again. It descends 1000 feet in the lower five-eighths of a mile, 500 feet of it in the last quarter mile and 200 to 300 feet in the terminal ice fall. The glacier gives an impression of ending on the rock lip of a hanging valley, though its terminus is probably on the face of a step in the main valley bottom. The western half mile of the glacier front ends in the bay, though in very shallow water, while the eastern quarter mile has a delta of outwash gravels and clay in front of it. This delta is an eighth of a mile wide at high tide and three-eighths of a mile at low tide. Across it flows one branching glacial stream from the eastern side of the glacier and a smaller stream from the hanging valley of Canyon Creek, which terminates in a lip a thousand feet above the fiord.

Rock ledges show beneath the glacier at one point behind the delta and at two points in the tidal front (Pl. C, A). Near the western edge the basal layers of exposed ice (Pl. XCIX) are dirt-laden and simulate rock ledges, though no rock was yet exposed there in 1909. This part of the ice front reaches tidewater, but icebergs are no longer discharged except by tumbling down the precipitous ice front, and the western half of the glacier is apparently just about to cease to be tidal. The water is so shallow that gravel talus cones are formed along the western edge of the ice margin.

[1] See Chart 8521 and 8519, U. S. Coast and Geod. Survey.

[2] Maps 20 and 21, Twentieth Ann. Rept., U. S. Geol. Survey, Part VII, 1900; and map accompanying Rept. XXV, War Dept., Adj.-Gen. Office, 1900.

Shoup Glacier is severely crevassed on this terminal fall but is little crevassed in the gentler slope above it (Pl. XCVIII, B) as far back as the ice fall at the elbow in the glacier. The front of the glacier is conspicuously free from débris-laden layers except in the basal portion just mentioned. On the surface there are no medial moraines and there is no lateral moraine on the eastern side, excepting below the elbow. The west side has a broad lateral moraine above the terminal ice fall, but this dirt and stone disappears in the crevasses so that except in the bottom ice there is practically no moraine on either margin at sea level.

Observations by Schrader and Grant. As already stated, the Shoup Glacier was not

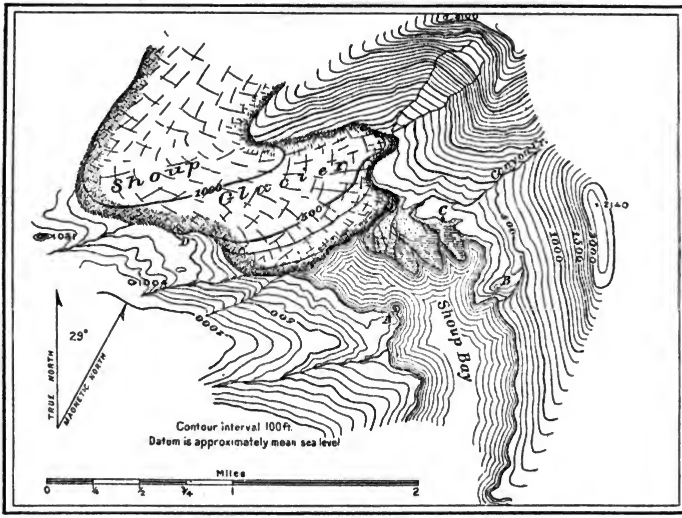


FIG. 24. SKETCH MAP OF SHOUP GLACIER IN 1909.

Owing to an accident to the instrument used there are slight errors in elevations and distances.

seen by Whidbey in 1794, and we know of no observations by the Russians who entered Port Valdez. F. C. Schrader visited and photographed it in 1898 and Emil Mahlo first showed it upon a map, but with the bifurcation already alluded to. Schrader briefly described the glacier in a U. S. Geological Survey report.[1] One of his photographs of the ice front, taken from a rocky knob on the west side of the bay (Photo. Station A), shows the exact conditions in 1898 and may be compared with photographs from the same site by the Coast Survey in 1901, by Grant in 1905 and 1908, and by the National Geographic Society's party in 1909 and 1910.

[1] See discussion of Canyon Creek Glacier valley in A Reconnaissance of a Part of Prince William Sound and the Copper River District, Alaska, in 1898, Twentieth Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 383–4.

In July, 1905, Shoup Glacier was visited by Grant whose descriptions, photographs, and maps [1] indicate that two large rocks were then being exposed by the retreat of the glacier that were not visible four years before. Along the sides of the glacier there was a broad space of bare ground free from soil and vegetation, and the whole aspect of the glacier indicated that it was retreating. On July 13, 1908, Grant and Higgins found that the front was in practically the same position as on July 4, 1905, as was also the case on June 16, 1909.

The statement by Grant that the two large rocks not visible in 1901 were being ex-

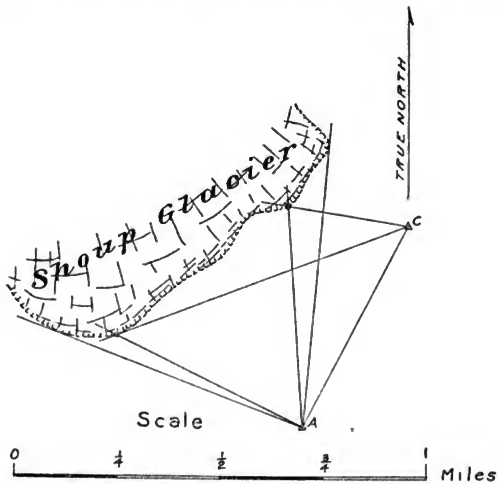


FIG. 25. TRIANGULATION ON FRONT OF SHOUP GLACIER.

posed in 1905, is of interest in view of the fact that in 1898, these ledges were already exposed, as shown by Schrader's photograph. Nearly five times as much of the larger rock showed in 1905 as in 1898. In 1908 the two areas of exposed rock ledges were nearly connected, the smaller or westernmost being of about the same size as in 1898 and 1905 while the larger, east, rock ledge was a little longer and higher in 1908 than in 1905. The progressive enlargement of the rock ledge areas between 1901 and 1905, and from 1905 to 1908, seems to indicate slow, though uninterrupted retreat during these years; but the fact that no ledges were visible in 1901, while two ledges were visible in 1898 and 1905 suggests a slight advance of Shoup Glacier between 1898 and 1901. Further evi-

[1] In H. F. Reid's Variations of Glaciers, Journ. Geol., Vol. XIV, 1906, p. 406; Vol. XVII, 1909, p. 670.
Grant, U. S. and Higgins D. F., Glaciers of the Northern Part of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLII, 1910, pp. 726–727.

dence of this advance is found in a comparison of the western margin of the glacier in the photographs taken from the same site in 1898 and 1905. The latter, by Grant, shows the ice edge a short distance farther west than in the 1898 picture by Schrader. This advance is also proved by other observations made in 1909, described below. Unfortunately Schrader's photograph does not extend far enough for us to compare the eastern margin of Shoup Glacier in 1898 and 1905. A comparison of Grant's 1905, 1908 and 1909 photographs of the two margins shows little, if any, retreat of the two edges.

Observations by the National Geographic Society's Expeditions. The National Geographic Society's expeditions spent August 19 and 20, 1909, and part of September 6, 1910, in a study of the Shoup Glacier, examinations being made of the eastern margin, the center of the ice front, and the southwest margin for some distance back from the glacier front. Photographs were taken from the site occupied by Schrader and Grant (Photo. Station A). The changes since Grant's last observations have been inappreciable. The topographic map reproduced as Fig. 24 was made in 1909 and bench marks were established on Photo. Station A and also on a point on the hillside to the west (Station C) from which transit readings were made as shown in Fig. 25. It is hoped that future observers may repeat the transit readings from these stations.

Transit Readings. Transit readings from Stations A and C, Shoup Glacier, August 20, 1909. Station A is an X on a rock on top of the first prominent spur south of the glacier on west side of Shoup Inlet at an elevation of 100 feet.

Readings

Point Sighted	*Ver. A*	*Ver. B*	*Angle from A–C*
1. Station C	180° 00′	00° 00′	
2. Most easterly point of ice front. (Same as No. 3, Sta. C)	148° 51′	328° 51′	31° 09′
3. Projecting point of glacier at first bend back of glacier front on east side of glacier near stream outlet	157° 48′	337° 48′	22° 12′
4. Southern point of glacier front on west side. (Same as No. 2, Sta. C)	89° 10′	269° 10′	90° 50′
5. Most western point of glacier	83° 56′	263° 56′	96° 04′

Station C is an X on a bowlder located on the east side of Shoup Inlet at an elevation of 130 feet above the fiord, and on the first large spur south of the glacier front close up to the edge of the thick alder.

Readings

Point Sighted	*Ver. A*	*Ver. B*	*Angle from A–C*
1. Station A	00° 00′	180° 00′	
2. Southern point of glacier on west side	41° 53′	221° 53′	41° 53′
3. Most easterly point of ice front visible from Sta. C	71° 52′	251° 52′	71° 52′

Former Expansion. There is proof of former expansion of Shoup Glacier from striae, roches moutonnées, etc.; but so far as we have evidence, the recent history has been one of retreat, with the exception of the slight advance mentioned above. Without knowledge of this recent forward swing, the fact of the advance was established in 1909 by the discovery of a recent moraine along the southwestern margin of the glacier. For the first half mile back from the shore, there was a minute morainic ridge winding up and down the hillside. It was from five to twenty feet from the glacier margin and was made up of angular fragments left by the receding glacier. There was wood in the débris and the stones at the edge lay upon dead shrubs whose neighbors were still growing. In some places there was a depression between the moraine and the glacier. Evidently a slight advance had interrupted the recession and the width of the barren zone indicated the amount of retreat since this recent advance.

During the more expanded ancient condition of the glacier, a group of barren rock hills farther north had been scraped clean of all soil and eroded into a series of roches moutonnées forms and rock basins, several of which contained lakes. The ice had retreated from these barren hills not long before the slight advance alluded to, but vegetation was finding a foothold slowly. Across this group of barren hills the small recent moraine extended in a sinuous course, and between it and the ice there was as yet almost no vegetation. Beyond these rock hills the area of recent overriding could be traced for several miles by the narrow barren zone between the little moraine and the glacier.

Evidence of the amount of time that has elapsed since the earlier, much more expanded position of Shoup Glacier is found in the distribution of forest on the neighboring hills. Willow, alder and cottonwood, mature, but perhaps not more than twenty or thirty years old, grow almost up to the very edge of the glacier. Scattered spruces are found within a mile or so of the glacier, but there is no mature coniferous forest at the ice front, as in the case of Columbia Glacier. Outside Shoup Bay, near sea level, scattered spruce extends up Valdez Fiord as far as Valdez, but higher on the mountain slopes none is found beyond the entrance of Shoup Bay. There is more spruce on the south than on north side of Port Valdez, and on the west than on the east side of Shoup Bay.

During the maximum of glaciation, when the extended Shoup Glacier received a tributary from Canyon Creek, the main Shoup ice stream over-deepened its valley so much that it left Canyon Creek hanging 1000 feet, though its erosion did not extend so far as to remove the ledges beneath the ice fall where the glacier front now rests; nor did it succeed in completely erasing the three, alternating, truncated spurs that project into the bay. When Shoup Glacier was tributary to the main Port Valdez Glacier, however, that larger ice tongue deeply eroded the main fiord, leaving the Shoup valley itself hanging 480 feet. This discordance, masked by the waters of the bay, is clearly shown in Fig. 26 where the greatest depth inside Shoup Bay is seen to be 45 fathoms, while the depth outside, in Port Valdez, 115 to 142 fathoms. There is a striking contrast between this submerged, hanging Shoup Glacier valley, the largest tributary of the great glacier that formerly extended down Port Valdez from the Valdez Glacier and Lowe River valleys, and the visibly hanging valleys of the smaller glacier tributaries, whose positions are indicated in Figure 26 by the letter h. In the mouth of each of these hanging valleys a steep-sided, post-glacial gorge has been cut, or a foaming waterfall descends, as from Canyon Creek. The different levels of the hanging valleys, and the highly perched Canyon Creek valley, hanging with reference to Shoup Bay, as that hangs with refer-

ence to Port Valdez Fiord, testify to differential ice erosion. The heights at which the valleys hang are roughly proportional to the size of the glaciers which occupied them and to the power of the glaciers whose erosion left them hanging.

Just after becoming detached from the Port Valdez Glacier, Shoup Glacier seems to have halted at the mouth of its bay, and to have spread out in a small piedmont bulb long enough to build a submarine deposit at the lip of its hanging valley; for here the

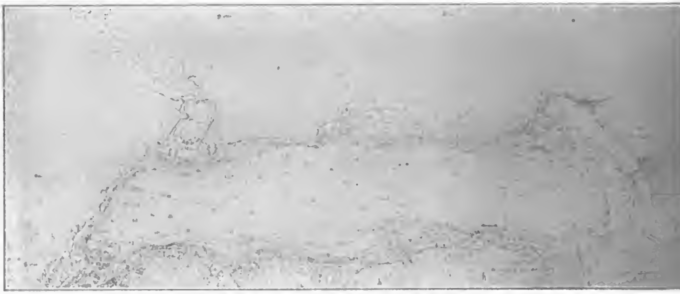


FIG. 26. SUBMARINE CONTOURS OF PORT VALDEZ.

After U. S. Coast and Geodetic Survey. Soundings in fathoms. Contour interval 100 feet. Hanging valleys shown by the letter h.

water now decreases in depth from forty-five fathoms within the bay to one or two fathoms just outside of it. This shoal at the bay mouth, and the sandspit upon it could hardly have been built in water 45 to 68 fathoms deep had not a morainic foundation been provided; and it is noteworthy that the deposit is just so located, and of just such shape, as a bulb glacier would build up here. After this period of halting Shoup Glacier seems to have retreated fairly steadily.

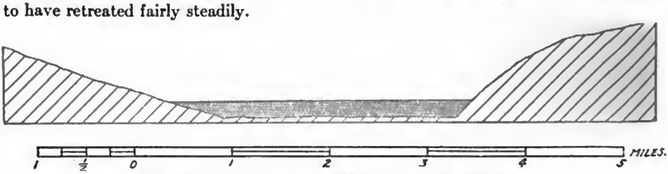


FIG. 27. NATURAL SCALE CROSS-SECTION OF PORT VALDEZ.

GLACIATION OF VALDEZ FIORD

It is evident that the entire Valdez Fiord has been at a former period deeply filled with a great glacier, tributary to the Prince William Sound glacier. We made no determinations of the exact height to which this glacier extended, but Grant and Higgins[1] estimate the height reached on the south side of Port Valdez as about 3200 feet.

As already stated, there are several visible hanging valleys on either side of the Port

[1] Grant, U. S. and Higgins, D. F., Bull. 443, U. S. Geol. Survey, 1910, p. 19.

Valdez Fiord, which testify to great glacial over-deepening of the main fiord. The detailed topographic map made by the U. S. Geological Survey in 1911 and the soundings in Port Valdez by the U. S. Coast and Geodetic Survey,[1] from which Fig. 26 has been made, show a number of features characteristic of a trough widened and deepened by glacial erosion. Above sea level the steepened lower slopes and the absence of spurs show the work of the former expanded ice tongue. Below sea level the trough characteristic of the fiord, a flat-bottomed, U-shaped form (Fig. 27), is exactly that of ice-eroded

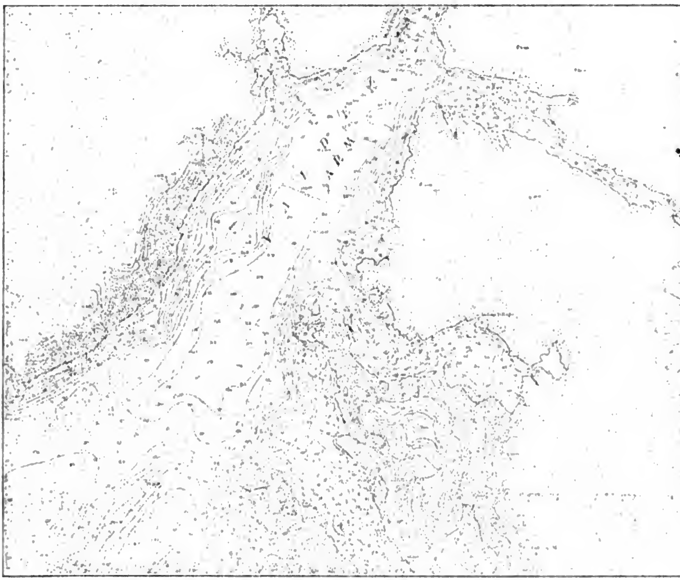


FIG. 28. SUBMARINE CONTOURS OF VALDEZ ARM.

After U. S. Coast and Geodetic Survey. Soundings in fathoms. Contour interval on the land 250 feet, and below sea level 100 feet.

valleys on the land. There are no known, uneroded lateral spurs below sea level. There are several submerged hanging valleys besides Shoup Bay, Sawmill Bay hanging 1150 feet (Fig. 28) and Galena Bay, north of Ellamar, 600 feet.

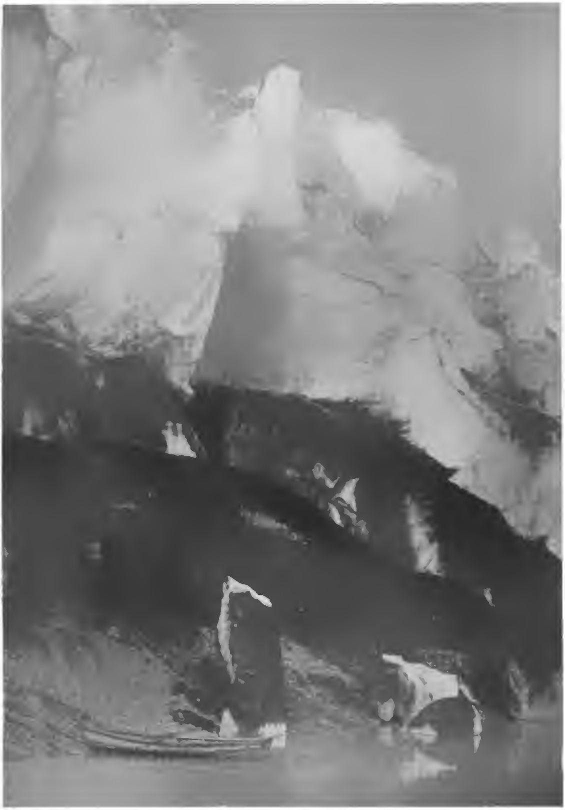
In the narrowest part of Valdez Narrows the fiord walls below sea level have slopes of 2500 feet to the mile, and even more in places. Here, where the ice stream was most

[1] Charts 8519 and 8521, U. S. Coast and Geod. Survey.

constricted, and might have moved fastest and eroded most efficiently there is a reef in midchannel (Middle Rock) which was not removed by glacial erosion, and which rises just above sea level at all stages of tide. The contours (Figs. 26 and 28) show that the part below sea level has the typical roches moutonnées form with the gentler slope on the side from which the ice came. South of it the fiord bottom is basined, probably by glacial erosion, and similar basining is also developed north of Valdez Narrows where the whole of the inner Port Valdez is 700 to 800 feet deep in contrast with 400 to 600 feet at the Narrows. Outside the Narrows, also, Valdez Arm seems to be basined, for the fiord is 1200 to 1250 feet deep but separated from outer Prince William Sound by a broad 200-foot barrier between Pt. Freemantle and Bligh Island. This may be either a submerged moraine or an uneroded reef. Galena and Jack Bays also have a basined character which may be explained either by bottom deposits or by differential glacial erosion.

So far as known there are no submerged moraines, except the one at the entrance to Shoup Bay. How much the fiord was filled by glacial deposits is not known, but the delta at the head of the fiord, built by the streams on the outwash plain in front of Valdez Glacier and by Lowe River, has doubtless encroached considerably upon the ice-eroded fiord and is still advancing at a rapid rate. Possibly also the flat floor of the fiord has been in part produced by glacial deposit. The fine mud and silt, shown on the Coast Survey chart, slopes only about 50 feet to the mile after leaving the steep descent of the submerged delta front near the town of Valdez.

Plate XCVIII



Shoup Glacier

A glacier front of only moderate height, but one beside which the man's figure (left of lower center) looks puny. Photograph, August 20, 1909.

A. Front of Valdez Glacier
Photograph by P. S. Hunt, from Station C (Fig. 22), June 21, 1909.

B. Shoup Glacier from Southwestern Margin
Photograph from Station D (Fig. 24), August 20, 1909.

Panoramic with left-hand picture.

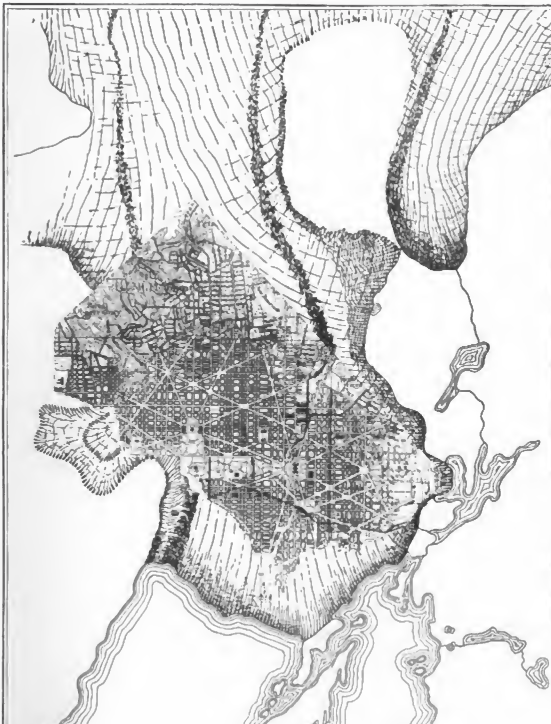
Panoramic with left-hand picture.

Plate C

A. Front of Shoup Glacier, August 20, 1909

B. General View of Four of the Cascading Glaciers of College Fiord
From right to left, Smith, Bryn Mawr, Vassar and Wellesley. Photograph from Station O (Map 7), College Point, July, 1910

Plate CI



Scale

1 ½ 0 1 2 3 MILES

The Lower Portion of Columbia Glacier
Showing the City of Washington, plotted on the same scale.

Plate CII

A.

B.

West Margin of Columbia Glacier

As photographed by Gilbert in 1899 (upper view) and by the National Geographic Society's expedition in 1909 (lower view), from Photo Station G, during the recent advance.

Plate CIII

Western Margin of Columbia Glacier
Encroaching upon the forest. August 23, 1909.

PLATE CIV

COLUMBIA GLACIER IN 1899

Steamer *George W. Elder* in front of the terminal cliff of Columbia Glacier in 1899. Photograph from "Harriman Alaska Expedition." Copyright, 1902, by E. H. Harriman.

PLATE CV

ICE TOWER AT THE FRONT OF COLUMBIA GLACIER

Which fell a few minutes after this photograph was taken, causing waves to wash up over 20 feet on the shore. Height of ice cliff compared with Bunker Hill monument, which is 220 feet high. Photograph, June, 1910.

A. FRONT OF COLUMBIA GLACIER IN 1899

On islet north of Heather Island. Photograph by G. K. Gilbert, from Station D (Map 6). See also Plates CVII and CVIII.

B. JUMBLE OF FOREST DÉBRIS

In front of the advancing Columbia Glacier on the islet north of Heather Island, June, 1910.

Plate CVII

A. Front of Columbia Glacier in 1905
On islet north of Heather Island. Photograph by Grant and Paige, July 10, 1905, from Station D.

B. Front of Columbia Glacier in 1908
On islet north of Heather Island. Photograph by Grant, July 15, 1908, from same site as upper view. See also Plates CVI, A, and CVIII.

A. FRONT OF COLUMBIA GLACIER, JUNE 24, 1909
On islet north of Heather Island. Photograph by Grant and Higgins, from Station D.

B. FRONT OF COLUMBIA GLACIER, AUGUST, 1909
On islet north of Heather Island. Photograph from a point close to Station D. See also Plates CVI, A, and CVII

Plate CIX

A. Columbia Glacier
Advancing and pushing up a terminal moraine, July 6, 1910.

B. Columbia Glacier
Advancing and pushing up a moraine and overriding forest on islet north of Heather Island, July 6, 1910.

A. Columbia Glacier in 1909

B. Columbia Glacier
Advancing into forest, overturning trees and shoving up a moraine, August, 1909.

A.

B.

C.

COLUMBIA GLACIER
On August 23, 1909, June 6, 1910, and September 5, 1910. Advance is rapidly narrowing the lagoon. From Photo Station N (Fig. 32).

A. TERMINAL MORAINE WITH KNOBS AND KETTLES
At eastern margin of Columbia Glacier. Photograph by U. S. Grant, June 24, 1909, from Station B (Map 6).

B. MEARES GLACIER AND MOUNT GROSVENOR
At the head of Unakwik Inlet. From Photo Station C (Fig. 37), July 13, 1910.

CHAPTER XIV

THE COLUMBIA GLACIER

Location and General Relationships. Glacier Island, which forms the western border of the entrance to Valdez Arm, lies directly in front of Columbia and Long Bays on the north shore of Prince William Sound. Columbia Bay is about three and a half miles wide at the entrance, but broadens to five miles inside, and contains Heather Island and several islets. At the head of the bay, five miles from the entrance, is the Columbia Glacier. In outer Columbia Bay the water is between nine hundred and a thousand feet deep, decreasing to six hundred feet near the glacier. The adjacent mountains rise from 2500 to 5000 feet. Columbia Bay is a broader and less imposing fiord than Port Valdez.

Explorations. The Columbia Glacier, which has also been called Live, Root, and Fremantle Glacier, was seen from the mouth of the bay and indicated roughly upon a map by Whidbey in 1794,[1] Applegate in 1887,[2] Mahlo in 1898[3] and Schrader in 1900.[4] Whidbey's map suggests that the glacier ended a short distance north of Heather Island in 1794 but does not distinguish the glacier from the land. Applegate's map represents the ice front in about the same position with respect to Heather Island. Mahlo's map does not show Heather Island. Schrader's map shows a small island, but like the other three does not represent the ice front with sufficient accuracy for comparison as to glacier movements. The only fact indicated by these maps is that the glacier was not strikingly different in the years of observation between 1794 and 1900.

Vancouver[5] tells of the conditions in 1794, when Whidbey saw Columbia Glacier from near the mouth of Long Bay, east of which was "another bay of rather large dimensions, with an island in its northeast corner . . . terminated by solid body of compact elevated ice, similar to that which has been before described . . . ; as they passed the eastern bay they again heard the thunder-like noise, and found that it had been produced by the falling of the large pieces of ice that appeared to have been very recently separated from the mass extending in vast abundance across the passage . . . insomuch that it was with great difficulty the boats could effect a passage."

[1] Vancouver, Capt. George, A Voyage of Discovery to the North Pacific Ocean and Round the World, 1790–1795, Vol. V, London, 1801, pp. 316-17.

[2] See map entitled Glaciers No. XI, in Davidson's The Glaciers of Alaska that are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc., Pacific, Vol. III, 1904.

[3] Map in pocket in book entitled Reports of Explorations in the Territory of Alaska, War Dept., Adj.-Gen. Office, No. XXV, Washington, 1899; Map No. 8 in Maps and Descriptions of Routes of Exploration in Alaska in 1898, special publication of U. S. Geol. Survey, 1899; Map No. 19, in Twentieth Ann. Rept., U. S. Geol. Survey, 1900, Part VII.

[4] Plate III in The Geology and Mineral Resources of a Portion of the Copper River District, Alaska, House Doc. 546, 56th Congress, 2nd Session, Washington, 1901.

One or another of these sketches of Columbia Glacier is also followed in U. S. Coast and Geod. Survey, Charts 8502 and 8519 and in Pl. I, Bull. 327, U. S. Geol. Survey, 1907.

[5] Op. cit., pp. 316–317.

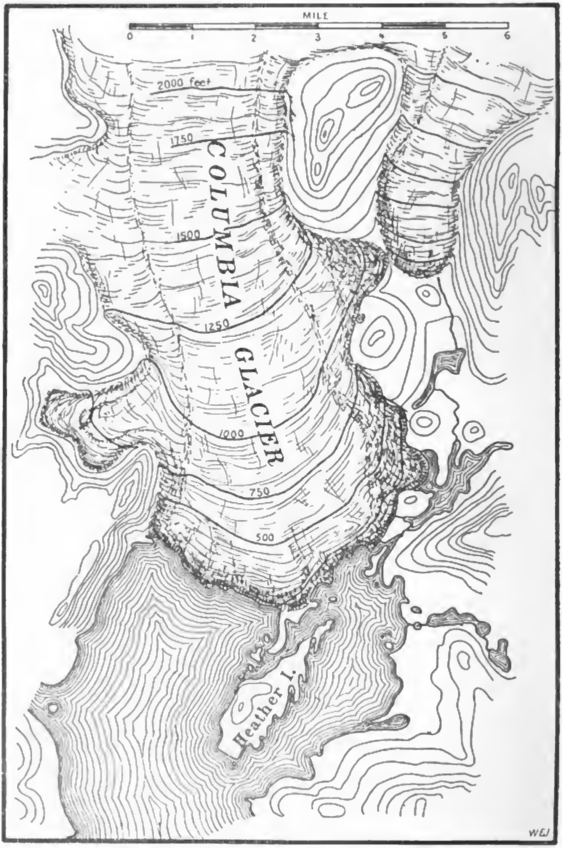


FIG. 29. GILBERT'S MAP OF COLUMBIA GLACIER IN 1899.

As stated in the preceding chapter Petroff's account,[1] thought to be from observations by some of the Russians and describing a great ice front in Port Valdez, may refer to the Columbia instead of the Valdez Glacier, for he mentions glaciers in Port Wells and Port Fidalgo but none in Columbia Bay.

Columbia Glacier was visited in 1898 by Capt. A. O. Johansen in the steamship *Dora* and a sounding of 50 fathoms was made near the ice front.

In 1899 the Harriman Expedition visited Columbia Bay and named and described the glacier,[2] Dr. G. K. Gilbert stayed there from June 25 to 28, studying the glacier, making the map here reproduced, and taking a number of photographs which have had great value in subsequent studies. His is the first and best scientific description of this ice tongue.[3] He found evidence of an advance in 1892. The proofs of this, and Gilbert's description of conditions in 1899, will be referred to repeatedly in the following pages.

In 1902 the steamship *Bertha* went within 300 yards of the ice front, and other steamships have visited Columbia Glacier in recent years.[4]

In 1905 a U. S. Geological Survey expedition in charge of Prof. U. S. Grant visited Columbia Glacier, and photographs were taken from the sites of some of Gilbert's photographs in 1899. Other photographs were taken from the same sites by Grant and Higgins in 1908 and in June, 1909, when the glacier was studied and the ice front mapped. The full report upon this work has not yet been published [5] but the abstracts [6] and photographs show that they observed a continuation of retreat here between 1899 and 1905, a slight advance between 1905 and 1908, and a greater advance in 1908 and 1909.

The National Geographic Society's expeditions have made four visits to Columbia Glacier, August 21 to 25, 1909, June 29 to July 11, 1910, September 5 to 6, 1910, and June 21, 1911. During 1909 and 1910 the glacier was still advancing, as evidenced by the deep thundering noise of the straining glacier, as well as by the lighter crash of ice masses sliding down the terminus and by changes along the terminus and margins of the glacier. We took photographs from several of the stations occupied by Gilbert, Curtis (photographer of the Harriman Expedition) and Grant, and observed a marked continuation of the advance seen by Grant, besides making soundings throughout Columbia Bay.

Our preliminary descriptions of Columbia Glacier have been published as follows:[7]

Our topographer, Mr. Lewis, has made the large scale contour map reproduced as

[1] Petroff, Ivan, Tenth Census of the United States, 1880, Vol. VIII, 1884, p. 27.

[2] Burroughs, John, Harriman Alaska Expedition, Vol. I, 1901, pp. 66–68.

[3] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, Glaciers and Glaciation, 1904, pp. 71–81.

[4] Higginson, Ella, Alaska, The Great Country, New York, 1908, pp. 257–259.

Greely, A. W., Handbook of Alaska, New York, 1909, pp. 156–158.

[5] Grant, U. S., Tidewater Glaciers of Prince William Sound and Kenai Peninsula, Bull. U. S. Geol. Survey. (In preparation.)

[6] In H. F. Reid's Variations of Glaciers, Journ. Geol., Vol. XIV, 1906, pp. 406–7; Vol. XVII, 1909, p. 670.

Grant, U. S. and Higgins, D. F., Glaciers of the Northern Part of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLII, 1910, pp. 727–735.

[7] Tarr, R. S. and Martin, Lawrence, National Geographic Society's Alaskan Expedition of 1909, Nat. Geog. Mag., Vol. XXI, 1910, pp. 9–10, 13, 30–33, 37, 53.

Martin, Lawrence, The National Geographic Society Researches in Alaska, Nat. Geog. Mag., Vol. XXII, 1911, pp. 548–553; Columbia Glacier, Alaska's Typical Ice Tongue, American Review of Reviews, Vol. XLIV, 1911, pp. 69–75; Two Glaciers in Alaska, Bull. Geol. Soc. Amer., Vol. 22, 1911, p. 731.

Map 6, as well as two sets of transit readings upon the ice front from stations listed below, which are readily accessible and should be revisited in years to come to determine the behavior of the glacier, whose advance from 1909 and 1910 is shown by Fig. 30, based

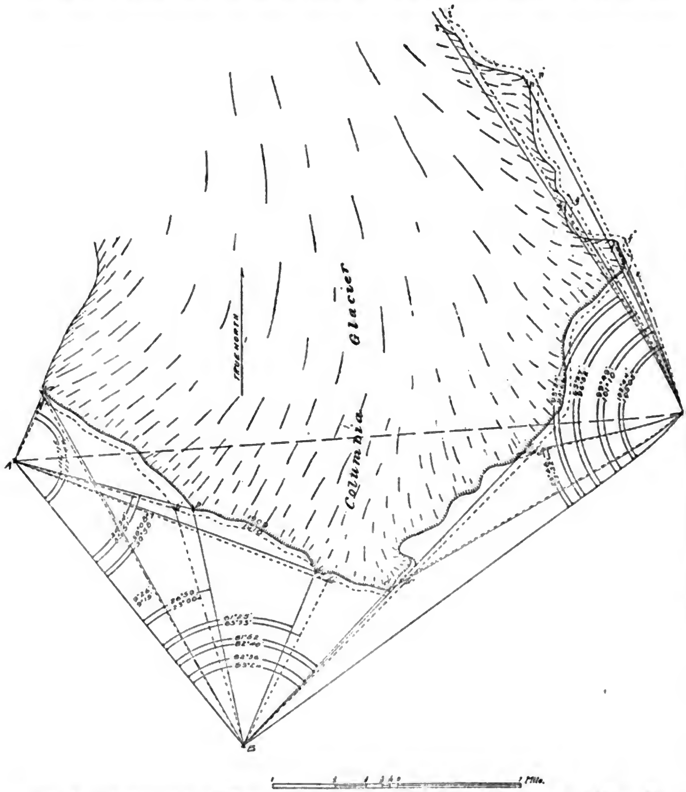


FIG. 30. TRIANGULATION ON FRONT OF COLUMBIA GLACIER SHOWING ADVANCE FROM 1909 TO 1910.

upon these transit readings. Stations B and C are close to the plane table stations occupied by Gilbert in 1899, and are each within a few feet of photographic stations H and I, see Map 6.

Transit Readings. Transit readings from Stations A, B and C, Columbia Glacier, August 23–24, 1909, and July 5–11, 1910.

Station A is located on the western side of Columbia Bay on a rocky spur about one-half mile south of the glacier front. It is a stake driven in the ground two feet east of twin trees, both of which are blazed on the east side. This station is close to but not the same as photographic station G, see Map 6.

READINGS

Point Sighted	*Ver. A*	*Ver. B*	*Angle from A–B*
1. Station B (Heather Island)	1909 0° 0′	180° 0′	
	1910 0° 0′	180° 0′	
2. Point on Heather Island where glacier enters timber	1909 329° 02′	149° 02′	30° 58′ (C)
	1910 329° 52′	149° 52′	30° 08′ (C)
3. Projecting point of glacier about half way between A and B.	1909 324° 09′	144° 09′	35° 51′ (B)
	1910 325° 26′	145° 26′	34° 34′ (B)
4. Most western point of glacier visible from Heather Island	1909 242° 12′	62° 12′	117° 48′ (A)
	1910 243° 45′	63° 45′	116° 15′ (A)
5. Where glacier enters forest on west side	1910 239° 47′	59° 47′	120° 13′
6. Point where glacier turns around the west spur	1910 245° 44′	65° 44′	114° 16′

Stat io B is on the highest point of Heather Island, 357 feet above sea level, and near the western side of the island. It is a stake driven in the ground and easily to be found.

READINGS

Point Sighted	*Ver. A*	*Ver. B*	*Angle from A–B*
1. Station A (West side of Columbia Bay)	1909 0° 0′	180° 0′	
	1910 0° 0′	180° 0′	
2. Most western point of glacier visible. (Same as No. 4, Sta. A.)	1909 9° 26′	189° 26′	9° 26′ (A)
	1910 9° 15′	189° 15′	9° 15′ (A)
3. Projecting point of glacier about half way between A and B. (Same as No. 3, Sta. A.)	1909 26° 59′	206° 59′	26° 59′ (B)
	1910 23° 00′	203° 00′	23° 00′ (B)
4. Point on Heather Island where glacier enters timber. (Same as No. 2, Sta. A.)	1909 61° 25′	241° 25′	61° 25′ (C)
	1910 65° 33′	245° 33′	65° 33′ (C)
5. Projecting point of glacier just east of Heather Island.	1909 81° 52′	261° 52′	81° 52′ (D)
	1910 82° 40′	262° 40′	82° 40′
6. Point where glacier leaves land on east side	1909 82° 36′	262° 36′	82° 36′ (E)
	1910 83° 24′	263° 24′	83° 24′ (E)

Station C, elevation 1497 feet above tide, at timber line on a long east-west spur east of the glacier, is a stake driven into a joint in a rock ledge.

READINGS

Point Sighted	*Ver. A*	*Ver. B*	*Angle from B–C*
1. Station B (Heather Island)	1909 0° 0′	180° 0′	
	1910 0° 0′	180° 0′	
2. Point of glacier just east of Heather	1909 6° 36′	186° 36′	6° 36′ (D)
Island. (Same as No. 5, Sta. B.)	1910 6° 58′	186° 58′	6° 58′ (D)
3. Point where glacier leaves land on east	1909 23° 46′	203° 46′	23° 46′ (E)
side. (Same as No. 6, Sta. B.)	1910 23° 32′	203° 32′	23° 32′ (E)
4. Most eastern projection of glacier toward	1909 102° 44′	282° 44′	102° 44′ (F)
inside lake.	1910 105° 05′	285° 05′	105° 05′ (F)
5. First bend in glacier north of No. 4.	1909 95° 27′	275° 27′	95° 27′ (G)
	1910 96° 52′	276° 52′	96° 52′ (G)
6. Next bend in glacier north of No. 5.	1909 99° 49′	279° 49′	99° 49′ (Ga)
	1910 101° 10′	281° 10′	101° 10′ (Ga)
7. Edge of glacier on west side of first nuna-	1909 93° 24′	273° 24′	93° 24′ (H)
tak.	1910 94° 26′	274° 26′	94° 26′ (H)
8. South edge of glacier on east side of sec-	1909 118° 00′	298° 00′	118° 00′ (I)
ond nunatak.	1910 118° 56′	298° 56′	118° 56′ (I)
9. West edge of glacier near the north end	1909 124° 36′	304° 36′	124° 36′ (J)
and on the east side of second nunatak	1910 125° 55′	305° 55′	125° 55′
10. Tongue of east branch of glacier north of alluvial fan.	1910 119° 56′	299° 56′	119° 56′

The points located by these two sets of transit readings when plotted (Fig. 30) show the advance of the glacier during ten months from 1909 to 1910.

General Features of Columbia Glacier. This giant among the glaciers of Prince William Sound has its source upon the southern slopes of the Chugach Mountains. Its length is not known, only the southern fifteen miles having been seen by white men. It rises near the base of Florence Peak[1], which is 11,190 feet high (Pl. XCIII) and is the highest point in the Chugach Mountains. It is possible that the length of Columbia Glacier is twenty-five miles or more and that it heads upon snowfield passes whence it is continued as a through glacier with the Valdez and Shoup Glaciers of Port Valdez to the east and the Tazlina Glacier of the Copper River drainage to the north. The lower part of the glacier varies in width from three to four miles, and the lower valley walls rise 2500 to 3200 feet.

The ice stream flows southwestward from its unknown source, and about nine miles from the sea turns abruptly southward. At the turn there is a 3664 foot peak, east of which the glacier sends a distributary southward for three and a half miles. This is one to two miles wide and at its terminus nearly, if not quite, coalesces with the main glacier,

[1] Named by the junior author of this book in 1910 for his wife.

making the hill a nunatak. At least one great south-flowing tributary enters the main glacier above this distributary, and there may be two other north or northwest tributaries. The west side has two pronounced embayments, the northernmost of which may contain a tributary glacier while the southern seems from one point of view to be simply a lateral valley into the mouth of which the Columbia Glacier margin has bulged westward.

The broad lower portion of Columbia Glacier is one of the largest and most beautiful in Alaska, terminating in a superb ice cliff from which icebergs are continually being discharged. The order of magnitude of this ice tongue may be seen by examining Pl. CI in which the whole city of Washington, D. C., has been drawn carefully to scale upon the surface of Columbia Glacier. One who has walked much in Washington can appreciate the magnitude of a surface of snowy-white, severely-crevassed ice twice as wide as from the Capitol to the White House and with a terminal ice cascade nearly as long as from the Navy Yard to the Naval Observatory, and varying in height from 100 to 250 feet.

There are medial moraines near each side of the lower glacier and a broad lateral moraine along the eastern margin, which is lobate and determined in position by several rock hills 500 to 1200 feet high. The distributary east of the nunatak is fronted by a low rock ridge and has its terminus covered with ablation moraine. The west lateral moraine of this distributary swings out into the ice tongue and becomes medial in the upper main glacier. The glacier is moving rapidly and is severely crevassed from side to side, though less broken within the area of lateral moraines.

For purposes of description Columbia Glacier may be divided into (a) the main glacier surface, (b) the lobate eastern margin, (c) the eastern ice cliff, (d) the Heather Island terminus, (e) the main ice cliff, and (f) the western margin. These will be described in inverse order.

The Western Margin. Gilbert observed in 1899 [1] that "at the western margin of the main ice cliff, where the glacier crowded against a steep slope, there was a belt of bare rock, from 200 to 300 feet broad, between the ice and the forest." A photograph taken by him from a rocky point about half a mile south of the western edge of the glacier front shows the conditions in 1899 and ten years later the National Geographic Society's expedition was able to reoccupy this exact site (G. Fig. 31), as Grant had done earlier in the same year. Our 1909 photograph is reproduced with that of Gilbert in Pl. CII and shows the precise change that had taken place in the ten year interval. There was considerable lateral spreading and thickening, for, as the glacier advanced, it had broadened so that the ice margin had pushed forward two or three hundred feet and once more reached the forest, into which it had not yet advanced very far on August 23, 1909. This is shown by the presence, in both pictures, of the same distinctive lone tree and the adjacent trees at the top of the hill between the ice edge and the forest. The glacier had thickened one hundred and fifty feet or thereabouts. The lateral spreading may have been more than 200 feet, for the glacier surely continued to recede between 1899 and the time when the 1909 advance began and it must, therefore, have readvanced this distance in addition to the 200 feet.

On July 5, 1910, the junior author found that this lateral spread had continued, covering the site of a marginal stream near the glacier terminus and crowding still farther westward into the forest. The amount of spread in the ten month interval was about

[1] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, Glaciers and Glaciation, 1904, pp. 76–77.

170 feet, trees being overturned and rocks overridden by the spreading ice. There was more wood than dirt and stone in the marginal débris heaped up by the spreading ice as it advanced into the forest. From Gilbert's photographic site on the rocky point to the south (G. Map 6) the evidence of the spreading was clearly seen and it was also plain that thickening had continued from 1909 to 1910, the amount being estimated as at least 50 to 75 feet.

Besides this spreading there was forward movement at the western margin from 1899 to 1910, for Gilbert says that in 1899 the belt of bare rock between the western margin of the glacier and the forest "was strewn with fragments not only of rock but also of wood, and trees were freshly overthrown at the margin of the forest. At the time of its attack on the forest the ice must have been 100 feet deeper than in the summer of 1899, and it also extended farther southward, as shown by a push moraine of rock at the water margin, 800 feet from the ice front. A second push moraine, less massive than the first, lay within it, being 160 feet from it at the water margin and elsewhere nearer to it than to the ice." He shows later, by the evidence of vegetation within this push-moraine along another part of the glacier, that this earlier advance had not taken place more recently than 1892 and perhaps in that year.

In June, 1909, Grant and Higgins found that the western margin had spread laterally far enough to encroach upon the forest and had advanced southward a distance estimated as about 500 feet beyond the position in 1899.[1] This advance had taken place mainly since Grant's earlier visit in July, 1908.

In August, 1909, we found that the western margin of the glacier had not only moved forward the eight hundred feet from the 1899 terminus to the outer push moraine, but that the front had also moved forward approximately four hundred feet more. The advance had without doubt been even more than the twelve hundred feet from the 1899 to the 1909 terminus, for the glacier surely continued to retreat between 1899 and the period between 1905 and 1908, when the present advance began, judging by its behavior on Heather Island during this period. We cannot determine how much this recession amounted to, but whatever it was must be added to the 1200 feet of net advance between 1899 and 1909. If Grant's estimate of the position in June, 1909, is correct, the advance between June and August, 1909, was about 700 feet.

In overriding the barren area strewn with rock and wood on the western margin, the two push moraines, mentioned by Gilbert, had been destroyed but a new push moraine had been formed in 1909, rising ten or fifteen feet above a beach where there was no such feature ten years before. It was made up of beach rubble and the ice was still in active contact with it on August 23, although the moraine was not visibly shoved while we were upon it. Above it rose the débris-stained, crevasse-riven ice cliff, from whose slope large blocks of ice had recently fallen upon the surface of the push moraine. The surface of this moraine was five or six feet wide and nearly flat-topped. It had a steep outer slope upon which several young spruces were still growing, having been tilted from a vertical position to an angle of forty-five degrees (Pl. XCVII, B), and in some cases even to a horizontal position, by the glacier pushing up the ground in which they grew.

The push moraine disappeared a few hundred feet back from the beach where the glacier was actively advancing into the forest along the western margin. Here we found a

[1] Grant, U. S. and Higgins, D. F., Glaciers of the Northern Part of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLII, 1910, p. 729.

jagged cliff, with ice dark from included rock fragments, in immediate contact with the steep mountain slope and the trees growing upon it. The trees were being uprooted and overturned by the advancing glacier (Pl. CIII) and their dead trunks were inclined at various angles. The trees not yet attacked were growing undisturbed, several of them beneath the very shadow of the overhanging ice cliff. The only attack upon the trees not yet reached by the ice itself was being made by the glacial stream which flowed along the ice margin. This stream had been forced westward into the forest and for some distance was flowing among the tree trunks, not having yet eroded sufficiently, or, at the base of the slope, deposited enough gravel to overturn or kill more than a very few of them.

On July 5, 1910, when the junior author revisited the western margin of Columbia Glacier, he found that the ice had not only spread and thickened since 1909 but had continued to advance southward during the ten months since his last visit. The amount of forward movement was about 340 feet and the terminus of the glacier was within 240 feet of the nearest rock ledges on the first promontory to the south, from which future measurements might conveniently be made, or from the bench mark near Station G (Map 6), still farther south.

The advance had taken the ice front over half way across the beach nearest the glacier and the marginal stream which emerged on the beach at the ice front in 1909 was gone, the water evidently escaping beneath the ice. The substantial push moraine of August, 1909, had also been destroyed by the advance and there was not as large an accumulation at the margin of the glacier. The ice was broken and splintered and upper portions were being thrust-faulted forward over the basal ice. In some cases trees were being pushed down by these upper ice splinters before the base of the ice cliff reached their roots.

Between July and September, 1910, this western margin of the glacier seemed from Heather Island to have advanced less than 120 feet. Between the latter date and June, 1911, the advance continued but as we saw the glacier only from the steamer, near the southern end of Heather Island, we cannot say how much except that it was less than 100 feet.

To summarize, the western margin of Columbia Glacier has had the following oscillations, the net advance from before 1909 to 1911 being at least 1700 feet. It is possible, in view of minor oscillations in midglacier that there were several short periods of retreat not mentioned in this table, and that the total advance was even greater.

Date of Change	*Nature of Change*	*Amount*	*Observer*
1892 to July 25–27, 1899	Retreat[1]	800 feet	Gilbert
1899 (July 25–27) to unknown date	Retreat	Not measured	
Unknown date to June 24, 1909	Advance	500 + feet[2]	Grant and Higgins
1909 (June 24) to Aug. 23, 1909	Advance	700 feet	Tarr and Martin
1909 (Aug. 23) to July 5, 1910	Advance	340 feet	Martin
1910 (July 5) to Sept. 5, 1910	Advance	100–120 feet	Martin
1910 (Sept. 5) to June 21, 1911	Advance	Less than 100 feet	Tarr and Martin

[1] Perhaps interrupted by minor advance about 1897 or 1898.

[2] Plus unmeasured retreat above.

The Main Ice Cliff. The main ice cliff of Columbia Glacier, between the west shore and Heather Island, has a width of a little over two and a half miles. It is somewhat less impressive than the Hubbard Glacier cliff in Yakutat Bay, being shorter, less active, and with less imposing mountains behind. It is, however, a beautiful, pinnacled (Pl. CV) and crevassed, snowy white, sinuous precipice of ice, rising vertically between 100 and 250 feet above the water, and in a less precipitous cascade to 500 feet. Its height, sweep and beauty may be inferred from Pl. CIV where the dimensions of the ocean steamship in the foreground give a scale for visualizing this magnificent ice cliff. It descends to an unknown depth below sea level, the nearest sounding being 600 feet, though the water may be a little deeper at other points. For various reasons, notably the sloping front of the ice cliff, the absence of large icebergs, and the shallowness of the water, we are convinced that the glacier front is not afloat. If it were, the water would need to be at least 1500 feet deep.

In 1899 Gilbert stated[1] that "at the western margin of its principal tidal cliff the glacier rested on a bank of drift at the level of low tide, and this bank extended eastward as a shoal,[2] on which bergs were stranded, for several hundred yards from the shore. We noted two important streams from the western ice cliff. One of them issued from a cave at the water's edge near the western limit of the cliff, the other from a submerged and invisible tunnel near the middle of the cliff. The last mentioned was probably the largest of all the draining streams. It rose to the surface at the base of the cliff and flowed southward over the salt water, forming a broad lane of milky fresh water with a visible current and at times nearly free from floating ice."

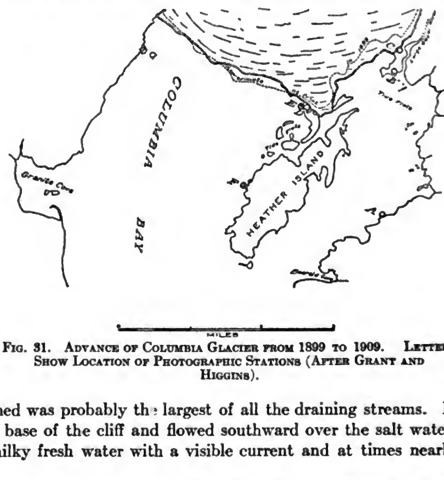


FIG. 31. ADVANCE OF COLUMBIA GLACIER FROM 1899 TO 1909. LETTERS SHOW LOCATION OF PHOTOGRAPHIC STATIONS (AFTER GRANT AND HIGGINS).

In 1909 and 1910 the streams mentioned were no longer there, doubtless having been destroyed by closing of orifices in the ice during the advance then in progress. The site of the shoal mentioned was overridden and there was no stranding of icebergs near the western margin except as ice masses had slid down from the cliff and lay upon the beach or near it in shallow water.

Grant and Higgins' map of Columbia ice front (Fig. 31), made in June, 1909, shows a retreat of parts of the west cliff of nearly one-eighth of a mile, since 1899,

[1] Op. cit., pp. 75–76.

[2] See Fig. 29. This was shown on his plane table map, but omitted in the reproduction in the Harriman Expedition report, Vol. III, Pl. XI.

there being an advance during the same period of ten years in the central part of the tidal cliff.[1]

In 1899 the point of the ice cliff projecting farthest south was near the center of the main cliff and it was in about the same position in August, 1909, though a little farther west. There were numerous minor promontories and coves in the ice cliff. The cliff was almost entirely free from débris, except in the areas reached by the western lateral moraine and one medial moraine near the western border. It was snowy white near the top and more glassy near the base, where there were also occasional minor dirt bands, also suggesting the shallowness of the bay here. Photographs, taken in 1909, from two sites on the small island north of Heather Island that were occupied by members of the Harriman Expedition ten years before, showed a slight advance of the main ice front in Columbia Bay and an increase in the height of the cliff.

With the advance of the cliff on the western margin there was a great increase in the height of effective wave work there, resulting in the destruction of vegetation by the great waves generated by the discharge of icebergs from the front of the glacier. Along the inner face of the beach and at the base of the cliff, vegetation had grown with relation to the height of wave work appropriate to an ice cliff a quarter of a mile or more away. When the ice cliff advanced this distance the waves began to wash up much higher, and in August, 1909, beach cobbles and sand had been thrown back among the trees, from some of which the bark had been eroded several inches above the ground. Some trees were killed by the sand and salt water, and some were broken off. Lichens and moss were removed from the rocky cliffs fifteen to twenty-five feet above high tide, soil had been washed down, and at the head of one chasm on the coast a wave had recently splashed so high out that it had ripped up a strip of turf and flung it, bottomside up, into the forest at the head of the chasm.

The main ice cliff, which descends nearly 500 feet in the last quarter of a mile, though frequently discharging icebergs, more by cascading down the front than by rising from below the surface, was far less active than Hubbard Glacier in Yakutat Bay. The terminal cliff of Columbia Glacier is nearly 200 feet higher than that of Hubbard Glacier, but it is far less steep. Columbia Glacier has a broken front down which ice fragments cascade whereas Hubbard Glacier front is in places nearly, if not quite, vertical for 300 feet. There were no large icebergs, apparently because the ice front stands in shallow water, and there were only a moderate number of small bergs, perhaps because of the easy escape they have to the open waters of Prince William Sound.

Neither the main ice cliff nor the eastern ice cliff projects as far as the Heather Island terminus, perhaps partly because the latter is opposite the center of the glacier, but also in part on account of the more rapid recession in salt water than on the land. Because of this difference, the ice front of Columbia Glacier is sinuous, with moderate re-entrants in the two places where in contact with the salt water, and with a notable projection where resting on the island.

In June, 1910, the main ice cliff was from 400 to 600 feet farther south than in August, 1909, the greatest advance (Fig. 30) having come in the portion a little west of the middle. The rate of actual advance of that portion of the glacier ending in water cannot be determined because iceberg discharge causes the shortening of the glacier while it is advancing. The rate must be considerably in excess of two feet a day, however,

[1] Grant, U. S. and Higgins, D. F., Bull. Amer. Geog. Soc., Vol. XLII, 1910, Fig. 6, p. 728.

for despite the losses by melting and by calving of icebergs, the main cliff was 600 feet farther south on July 5, 1910, than on August 24, 1909,—314 days earlier.

Pictures from two photographic sites occupied by the Harriman Expedition in June, 1899, and by us in August, 1909, June, 1910, and September, 1910, enable us to check, with graphic proof, the forward movement determined with the transit (Fig. 30). The advance from August, 1909, to June, 1910, was much greater than that of the preceding ten years (Fig. 31). The advance from June, 1910, to September, 1910, which was not measured with the transit, was slight, though perceptible. It may be estimated roughly as over 100 feet.

During the continued advance from June to September, 1910, a large area of débris-laden, black ice appeared in the terminal cliff some distance east of the center. It was several hundred yards long and extended 50 or 75 feet above sea level.

The Heather Island Terminus. East of the main ice cliff the Columbia Glacier terminates for about half a mile on an islet north of Heather Island and connected with it at low tide (Fig. 35). The ice front ends successively from west to east (Fig. 32): (1) on a beach (a–c); (2) near a western clump of timber (c–d); (3) on a boggy heath (d–e); (4) near an eastern timber belt (e–h); and (5) on morainic shoals and mud flats in the cove north of Heather Island and east of this islet (h–i). Beyond this is the eastern ice cliff.

This part of the glacier has been visited and studied more than any other and the events in its history for the past ten years are of much interest. In 1899 Gilbert[1] states that "on the island between the two ice cliffs there were also two push moraines of recent date, the nearer being about 100 feet from the ice front, the farther from 300 to 500 feet. The latter was associated with overthrown forest trees, and included with its rocky débris not only tree trunks and branches but folds of peaty soil. The tract between the nearer push moraine and the ice was in places occupied by an old moraine surface over which the ice had advanced, but this surface was elaborately fluted in the direction of ice motion, the corrugations having a vertical magnitude of several feet. In one instance it was seen that a large bowlder in the underlying drift had impressed its form on the ice, preserving in its lee a train of drift of the same cross-section, which constituted a ridge, and it is probable that the other flutings were of the same character. As these details in the configuration of the drift surface would be quickly obliterated by frost and rain, their exposure must have been very recent. Probably the advance creating the push moraine and the subsequent melting which laid bare the ice-molded drift had taken place within one or two years."

Grant visited this same part of Columbia Glacier in 1905, 1908, and June, 1909, and has stated[2] that in 1905 "at the north end of the small island north of Heather Island, on which the front of the glacier is resting, and where a few years ago the glacier had intruded and overturned the front of a forest, a photograph (CVII, A) was obtained from the same position as one (Pl. CVI, A) taken by Mr. Gilbert in 1899 (Fig. 31, Sta. D). At this point the front of the ice has retreated 160 feet since 1899. On the ground since vacated by the glacier there is very little vegetation—practically nothing except fireweed, which has encroached upon this territory only a few feet."

[1] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 77–78.

[2] Grant, U. S., in H. F. Reid's Variations of Glaciers, Journ. Geol., Vol. XIV, 1906, pp. 406–7; Vol. XVII, 1909, p. 670.

"In 1905 the front had retreated 160 feet from its position in 1899, and in 1908 it had advanced 112 feet beyond its position in 1905."

In June, 1909, when revisited by Grant and Higgins this part of the front of the glacier had advanced 310 feet since 1908, and along parts of its front was advancing into the forest.[1]

The photographs by Gilbert in 1899 and by Grant in 1905, 1908 and June, 1909, from

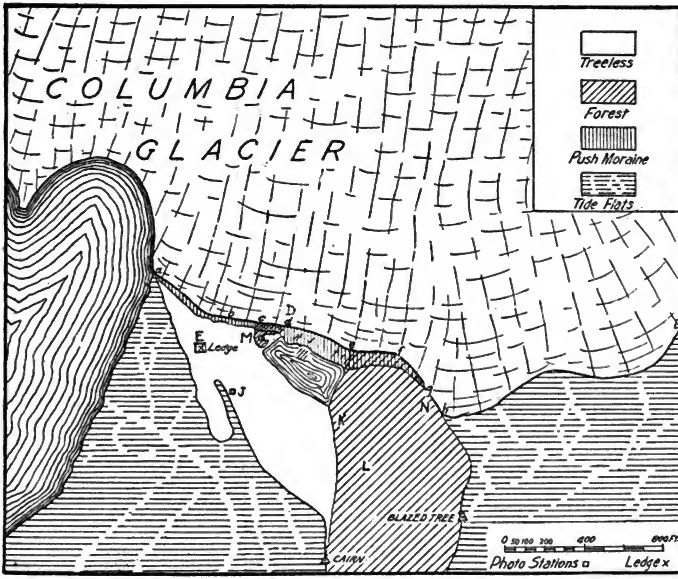


FIG. 32. THE HEATHER ISLAND TERMINUS OF COLUMBIA GLACIER IN JUNE, 1910.

Small letters show location of forest belts, heath, beach, etc. Capital letters show photographic stations, those not precisely located having a letter without symbol.

exactly the same site, at the western end of the eastern timber belt already spoken of, show the retreat and advance at this point graphically (Pls. CVI, A, CVII and CVIII). It is also a noteworthy fact that the advance went on even more rapidly between June and August, 1909, than before, so that while Grant was able to photograph the glacier from this site in the former month, the National Geographic Society's expedition, two months later, found the site completely buried beneath the glacier, which had then advanced

[1] Grant U. S. and Higgins, D. F., Glaciers of the Northern Part of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLII, 1910, pp. 731–733.

up to or possibly just beyond the outer push moraine observed by Gilbert in 1899. Our photograph (Pl. CVIII, B), which belongs with this remarkable series of four, was therefore necessarily taken from a little farther to the right and higher than the others. It shows, however, the same group of inclined and overturned, dead or dying trees which are in all the other pictures and which were killed by the advance dated by Gilbert as seven years before 1899, or in 1892. In August, 1909, the ice front had just passed this 1892 maximum. On July 4 and September 5, 1910, the advance had gone still farther, as is shown below. The measured record at this point is as follows.

Date of Observation	*Nature of Change*	*Amount*	*Observer*
1892 (probably) to July 25–7, 1899	Retreat	3–500 ft.	Gilbert
1899 (July 25–7) to July 10, 1905	Retreat	160 ft.	Grant
1905 (July 10) to July 15, 1908	Advance	112 ft.	Grant
1908 (July 15) to June 24, 1909	Advance	310 ft.	Grant and Higgins
1909 (June 24) to Aug. 23, 1909	Advance	70 ft.	Tarr and Martin
1909 (Aug. 23) to July 4, 1910	Advance	200 ft.	Martin
1910 (July 4) to Sept. 5, 1910	Advance	50 ft.	Martin

The photographs at these seven periods of observations, show many interesting details concerning the forms of retreating and advancing glacier fronts, the changes of marginal drainage, the accompanying push moraines, the effect on forests and the rapid reoccupation by annual plants of a deglaciated surface.

The half mile of ice front of Columbia Glacier on this islet north of Heather Island showed the following conditions on August 22–23, 1909, July 4–9, 1910, and September 5, 1910. Beginning where the main ice cliff emerged from the water of the bay on the western side, there was a push moraine of beach sand and gravel 15 or 20 feet high in August, 1909, where the photographs show that there was no such moraine in 1899, 1905, and 1908. The ice was in active contact with this. An older push moraine, probably built by the 1892 advance, still extended some distance out toward the water's edge in 1908 but it did not extend to the coast, as did that of 1909, which is farther south. A tiny push moraine within this older one also shows in Grant's 1908 picture, the advance having then apparently commenced. At the time of Grant's[1] visit in June 1909, the outer moraine had advanced and grown, being about 25 feet high. It was formed between July 15, 1908, and June 24, 1909, and continued to be pushed forward from June 24 to August 23, 1909. Between August 23, 1909, and July 4, 1910, the glacier margin upon the beach continued to advance, shoving the push moraine upon the beach ahead of it and truncating more and more of the forest edge.

Measuring from a graywacke rock ledge upon the beach which was about 300 feet from the ice in August, 1909, the glacier margin advanced 135 feet in the ten months up

[1] Grant, U. S., and Higgins, D. F., Glaciers of the Northern Part of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLII, 1910, p. 735.

to July 4, 1910. At the same point the advance from July 4 to September 5, 1910, two months, was 132 feet. At this latest observation, the edge of the push moraine in front of the advancing glacier was only 33 feet from the highest point in the middle of the graywacke ledge.

At the time of each of the three visits by the National Geographic Society's expeditions the push moraine at the advancing glacier margin showed about the same conditions. There was a gravel ridge parallel to the ice front and 15 to 25 feet high (Pl. CIX, A). Between its crest and the ice was a depression 6 or 8 feet deep, and the nose of the glacier was visibly plowing under the gravel. The outer slope of the push moraine was as steep as sliding gravel will stand. The advance was taking place without disturbance of the beach gravels beyond the push moraine, in this respect contrasting with the conditions along the eastern edge of the glacier which are described later.

The western edge of the western clump of timber, as shown in an 1899 photograph, was not recognizable in August, 1909, and at the time of our visit in 1909, there seemed to be no trees as near the water's edge as in 1899. Since the forest belt recedes farther from the coast as one proceeds southward from the 1899 glacier front this indicates a considerable advance before August, 1909, truncating the edge of the forest, and extending beyond the 1892 maximum. Within the western clump of timber the ice was pushing in among the trees in August, 1909, overturning them by the thrust against their roots and killing them by burying their trunks beneath the rock and soil of the push moraine which bordered parts of the ice front. Furthermore, the glacier was faulted in places and long splinters of ice were thrust forward in advance of the base of the ice cliff, coming in contact with the upper trunks of mature trees and tipping them over. These upper ice portions were being pushed ahead of those below by thrust faulting, evidently because the basal ice was in contact with the soil and the tangle of roots of a mature forest, so that the less resisted upper layers of the ice slid over those below. There were small living trees actually under the projecting ice front. One could step from the glacier surface into the branches of a tree twenty-five feet above the ground, or climb a tree and step out upon the surface of the projecting upper layer of ice.

The exact amount of advance here is not definitely known, but the ice was surely farther out than it had been for fifty to one hundred years, judging by the age of the trees now being overturned. The forest was mature, with thickly-set trunks and deep moss, and some of the trees were a foot in diameter and from fifty to one hundred years old. That at the time of our visit in August, 1909, the advance at this point had extended beyond the 1892 maximum is proved by the absence of previously overturned trees in this western forest belt, in contrast with the recognizable inclined trees just to the east where the advance had extended up to but probably not very far beyond the 1892 maximum.

Between August, 1909, and July 4, 1910 (Fig. 32), the ice advanced nearly through this western belt of timber, overriding the trees and pushing up a mass of soil, gravel, and wood in front of it. During this period of ten months the width of the forest from north to south was diminished from about 300 feet to 100 feet in the widest place. Between July 4 and September 5, 1910 (Fig. 33), the destruction of this forest continued and on the latter date there was not a tree left standing upright in the main grove. All were either overridden or lay prone (Pl. CIX, B), or were held inclined by the moraine or the ice, except a half-dozen semi-detached spruces at the southern border. One of

these was already inclined in September, 1910, suggesting that a ploughshare-like basal nose of the glacier was advancing ahead of the visible ice edge. In the intervening area a few rolls of turf were being pushed up and these increased in number during the two months between our 1910 observations.

In the forest there was no audible disturbance that would lead one to suspect that the glacier was advancing, and it was possible to clamber freely among the overturned tree trunks, over the push moraine, and out on the edge of the advancing glacier, whose ice was less splintered and overthrust in July and September, 1910, than in August, 1909. More of the trees seemed to be thrown down by the weight of the glacial moraine material heaped upon their trunks than by the push of the ice itself, therefore falling with their tops away from the ice and afterward being overridden and buried. Some of

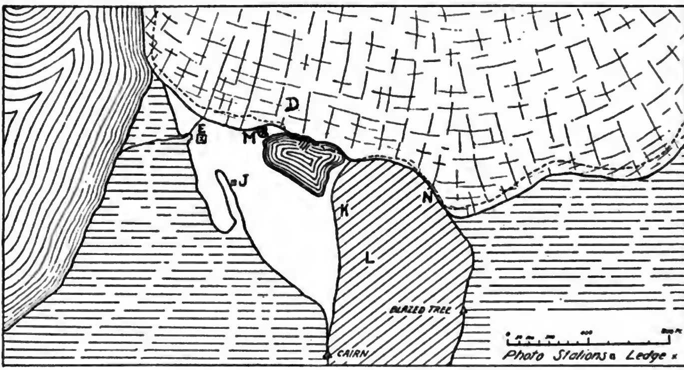


Fig. 33. Columbia Glacier in September, 1910.

To show advance during 2 months. Dotted line shows ice edge in June, 1910, see Fig. 32.

the trees, however, were originally overturned by the ice itself and these usually fell toward the glacier, the projecting, buried ice edge pushing their roots forward. They, therefore, lay across the terminal push moraine and in some cases with their tops on the ice itself. Ice seemed to everywhere underlie this terminal push moraine, though in places it was deeply buried.

The terminus of that part of the glacier which ends on the heath between the two forest belts was fronted in August, 1909, by a great push moraine of peat and turf with a few tree trunks, some bowlders, and a little gravel and till. This push moraine rose twenty to twenty-five feet above the heath on the south, which sloped gradually to the shores of a small pond. The ice-contact face of the push moraine sometimes had a depression between it and the ice which was due to the thrust, not to melting, for the grasses on the inner slope had not been destroyed. The shoving up of the push moraine was evidently due mainly to the powerful ploughing of the glacier foot and the thrusting forward of the soil.

This push moraine was in part a feature produced by the 1892 advance, for there were dead trees in it overturned by the previous advance, and a photograph taken in 1899 shows the moraine crossing the heath much as in 1909, though with a very different relation to the ice front. In August, 1909, the ice was just extending beyond the 1892 maximum, as was shown by fresh clods of turf, stones, and pieces of wood that had rolled down the outer slope of the push moraine and by one or two blocks of ice that had tumbled down the outer side. We are not certain whether the moraine here had been moved forward at all or had been made higher in 1909.

By September, 1910, there was still more advance, but the glacier terminus was much thinner. This spectacular decrease in height of the ice terminus from 60 feet in July to less than 15 feet in September suggests either (a) tremendous vertical ablation (45 feet removed in 63 days, or nearly nine inches in a day), or else (b) the collapse of this portion of the glacier surface by flowing away of the ice beneath.

In this eastern part of this heath or meadow the push moraine was fronted by an area of turf ridges or billows, extending parallel to the ice front and one hundred and fifty or two hundred feet out from it. In 1909 there were synclines and anticlines of peaty soil two to six feet wide and three or four feet high, with grass and shrubs still growing upon them. Some of them had been made by the thrust of the ice during the 1892 advance, for they were photographed in 1899. At least two such peaty bolsters were, however, added by the 1909 advance at this point. These are not rolls that were invisible because submerged in the adjacent pond in 1899, for the water stood higher in 1909 than ten years before. The advance at this point extended, therefore, a trifle farther in August, 1909, than the 1892 maximum.

In July, 1910, the push moraine was quite similar to that in August, 1909, except that the whole morainic ridge had been pushed forward and it continued to advance during the summer of 1910. The sloping heath in front of it, however, was replaced by additional bolsters of peat. The depression between the ice and the push moraine was decreased somewhat in depth between 1909 and 1910. The push moraine rose 10 or 12 feet above the ice of the glacier terminus, and 22 feet above the lake to the south. It was made up of till with enormous striated bowlders, rounded gravel, and peat. It was covered with grass except in places where cracks interrupted and upon it were a few trees, most of them dead or dying, as a result of the advance of 1892. These trees rode forward on the back of the moraine from 1909 to 1910, maintaining their relative positions but slanting at different angles. As already stated, a number of peat rolls were added during this period of advance, the whole heath which sloped forward in front of the moraine in 1909 being destroyed before September, 1910. In some places two or three peat rolls were added between July and September, 1910. Some of these arches of peat were six feet high, running parallel to the terminal moraine. In a few cases the pressure had resulted in closing up the folds to tightly-packed isoclinal structures, in others the arches had cracked across their axes, revealing the internal structure. The lake in front of this portion of the ice front was of about the same size in 1909 and 1910, but of different shape and of slightly greater depth when the 1910 observations were made.

At the eastern end of this heathy meadow was a group of overturned and inclined trees, some dead and a few living. These merged into the eastern timber belt. Through this grove extended a push moraine made up of an intricate jumble of till, gravel,

stones and tree trunks, the latter, some of which were over two feet in diameter, being inverted, prostrate, inclined, or beginning to be overturned. Against this moraine the ice cliff of the glacier extended closely in 1909 (Pl. CX, A), being at least as far out as the advance that killed the trees in 1892, and perhaps a little farther. The glacier terminated against this push moraine, above which it towered precipitously, but it was not so severely crevassed that one could not walk out a short distance upon its surface. At one point we could enter some of its larger crevasses and see the deep blue of the ice from within. The dirty basal ice projected farthest in this area, there being no projecting splinters of clear upper ice thrust forward among these trees, as there were in the western timber belt.

Outside the push moraine, in the tangle of tree trunks, there were a few rolls of peaty soil with shrubs and other plants, still growing in 1909, although often at right angles to their normal vertical position. One such roll was a typical anticline overturned on the side away from the thrust, whose effect here was felt at least 100 feet from the ice. Few of the growing trees which were not affected by the 1892 advance had been overturned in August, 1909.

At the eastern end of this forest belt a small pool was held in between the ice front and the rocky slope of the higher part of the island (Pl. CX, B), and its outlet flowed westward through the edge of the forest to the small pond near the heath mentioned above. Out of this east pool rose a series of overturned and compressed folds of peaty soil on the extreme eastern edge of the belt of overturned trees. On the opposite (east) side of this pool the glacier had pushed up a former sea bottom deposit of clay with abundant marine shells.

In 1910 the whole situation was changed, owing to the advance of the glacier. A few of the inclined trees were recognizable, but the ice front had advanced several hundred feet into the growing forest and had heaped up among the growing trees a terminal moraine mass of tree trunks, soil, rock, and vegetation (Pl. CVI, B). It rose to a height of 30 or 40 feet, at least, where ten months before there had been no surface accumulation. Above this heap rose the inclined trees in a confused thicket, with trunks pointing at every conceivable angle. The growing trees in this eastern timber belt had not been overridden and buried, as in the western timber belt, nor pushed forward without being overturned, as was the case with the scattered trees on the intermediate push moraine back of the heathy meadow. All of them were overwhelmed by the advancing glacier, but only a few were overridden, while most were heaped up in the windrow of forest débris, whose edge advanced steadily into the growing forest from July to September, 1910. Some distinctive blackened tree trunks, which had been killed previously, changed position from August, 1909, to July, 1910, not only by forward movement but by being lifted vertically 30 or 40 feet.

Along the eastern edge of this forest belt, in 1910, the site of the little pool observed in 1909 was overridden; and the glacier had advanced to the edge of a high bank which was a hundred feet or more from the ice the year before. As yet there were no push moraine or tangle of tree trunks at the margin of the glacier, which had advanced across a clear space formerly occupied by the eastern pool and by a marginal stream course. In July and September, 1910, however, the forest was just being invaded by the ice. One tree which was seven feet from the ice on July 6th was touched by the glacier on July 9th, the advance being 2⅓ feet a day; and before September 5th this tree was over-

ridden. Measurements made from the glacier edge to a blazed tree at the base of a neighboring hill showed an advance here of 20 feet from July 9th to September 5th, the rate being slower than in July, probably because the glacier was advancing up a steep slope during the latter period. On September 5th the ice edge was only 13 feet from the tree, which is blazed on the side away from the glacier.

The remainder of this portion of the ice edge was bordered in 1910 by a strip of mud, covering the terminus of the glacier. There was a series of mud flows of varying liquidity depending upon the amount of water from the melting ice, some of them flowing out into the borders of the forest, overwhelming the shrubs and rising up around the trunks of the trees. Near the eastern edge of the timber belt the mud flows were crowding in against a high bank, along which they rose appreciably higher in September than in July, 1910.

In the cove between the eastern end of Heather Island and the small island to the north of it the glacier ended in a low, dirty, sloping front instead of in a clean, precipitous cliff. The actual terminus was upon morainic shoals and tidal mud flats so that the water came up over the end of the glacier at high tide; but at low tide the cove had no water except a narrow shallow stream flowing down the middle. Near the glacier were many mud flows, and the whole cove, or lagoon, was floored by fine glacial mud across which it was possible to walk at low tide. Scattered through it were a few bowlders dropped by icebergs which floated in the cove at high tide, and upon its surface there were iceberg pits and long, sinuous trails left by the dragging of icebergs during the rising and falling tide. None of these icebergs were discharged from this part of the glacier but floated in from the larger bay to the east. Numerous small streams emerging from this part of the glacier were helping to fill the cove with glacial mud.

This cove was narrowed considerably by the advance of the glacier from August, 1909, to July, 1910, and from the latter date to September, 1910, as the three photographs of Pl. CXI clearly show.

That this part of the ice front extended farther out a number of years ago is indicated by a morainic island extending at right angles to the present glacier terminus and not now covered at high tide. It has no trees or other vegetation, except annual plants, and is the only land close to the ice front that has neither trees nor thick heather; but the annual plants that grow upon it, show that the tide does not rise over it now. Gilbert says [1] that in 1899 "a bank also extended eastward from the island against which the ice front rested, constituting at low water a stony cape half a mile long near the foot of the ice cliff." His description and photograph do not harmonize with our observations in 1909, for the bank was surely an island then. It is possible that this part of the glacier has advanced to the island since 1899, for no other cause for raising it above high tide is possible, excepting uplift. Recent advance to the morainic island is indicated by the presence of a sinuous ridge of gravelly material on the side next the ice front. This deposit may possibly represent either an esker or a push moraine. The latter seems the more probable, for in July, 1910, the advancing ice front was building a low push moraine with a series of minor lobes which had narrow crests and very angular turns, similar to those in the higher, sinuous deposit of older date.

In front of the ice there was a broad, low arch of stiff mud with marine shells, pushed up between August, 1909, and June, 1910, across which an antecedent stream had cut a

[1] Op. cit., p. 75.

narrow gully, deepest at the arch of the uplift, where there was a little cascade. The whole arch is 100 feet or so in width.

That icebergs were formerly discharged in this cove, when the glacier extended out farther than now, and the water was not yet so shallowed by deposition, is indicated by an abandoned wave-cut cliff on the Heather Island side of the lagoon. At the top of this cliff were trees fifty to seventy-five years old; but the face of the cliff is partly covered with a growth of peat, growing in which and at the base of the cliff were trees fully twenty-five years old. The cliff-cutting had, therefore, ceased over twenty-five years before 1909.

It is not known exactly what changes in detail took place in this part of the terminus of Columbia Glacier between 1899 and 1908, but between 1908 and late August, 1909, there was sufficient advance of the terminus so that the details shown in Grant's 1908 photographs were entirely destroyed by August, 1909. Marked thickening of the ice in this part of the terminus is demonstrated by a comparison of the Gilbert photograph from the islet north of Heather Island in 1899, with Grant's 1905 and 1908 photographs from the same site. These show that there was not only an advance and thickening between 1899 and 1905 but that the advance ceased, the ice thinned again by ablation and was again thickened by advance between 1905 and 1908, the thickening continuing very markedly in 1909 and 1910.

In Gilbert's 1899 photograph (Pl. CVI, A) the part of the ice front beyond the eastern timber belt shows as a low, irregular surface, completely mantled by débris. In Grant's 1905 photograph (Pl. CVII, A) this area is occupied by crevassed ice, at least fifty feet thicker than in 1899, judging by the height to which it rises upon the nearer inclined tree trunks through which it shows. It may be even thicker. Following this advance and thickening there had been some ablation, for the terminus was again partly mantled with débris and the crevasses partly healed. This advance had, therefore, ceased before 1905 and had not extended out any farther than the 1892 maximum, as shown by stones in the foreground of the 1899 and 1905 pictures and by Grant's measurements of the retreat of the ice front several hundred yards to the west. Comparison of Grant's 1905 and 1908 photographs show that the retreat continued some time after 1905, for the surface seen through the trees was lowered very considerably by ablation between 1905 and 1908, though still thicker than in 1899. The surface to the left of the trees in the 1908 photograph (Pl. CVII, B), shows a great crevassed dome of clear ice, indicating that a new advance was then in progress, this being also shown by the advance of the glacier margin as measured by Grant. The thickening between 1905 and 1908 cannot be determined closely, for we do not know how much lowering by ablation there was after 1905 and before the 1908 advance began; but it was evidently not less than fifty or seventy-five feet, and may have been much more. This 1908 advance continued into 1909 and 1910, being observed by Grant in June (Pl. CVIII, A) and by the National Geographic Society's expedition in August, 1909, and on our two visits in 1910.

Thickening and forward movement of this part of the terminus is also shown by comparing Gilbert's 1899 photograph from the 1497 foot station (I) on the hillside to the east with our own taken in August, 1909. Projecting the terminus in the two photographs forward against the islands north of Heather Island it is evident from Gilbert's map that the net advance in the last ten years is approximately five-sixteenths of a mile or about 1600 feet. It is not known how much of this was between 1899 and 1905

and how much between 1905 and 1909, nor how much should be added for readvance after continuation of the retreat that was in progress in 1899. Between August, 1909, and July, 1910, there was additional advance here of about 165 feet, followed by additional advance from July to September, 1910. As the advance on the western margin is at least 1700 feet an advance of over 1600 feet near the middle of the glacier during this decade is not abnormal. It suggests, of course, that there was similar greater retreat and readvance on the north of islet Heather Island and that the table based on Grant's measurements, should be amplified as follows:

Date of Observation	*Nature of Change*
Unknown date to 1892 (probably)	Advance
1892 (probably) to July 25–7, 1899	Retreat [1]
July 25–7, 1899, to unknown date	Retreat continued
Unknown date to second unknown date	Advance
Second unknown date to July 10, 1905	Retreat
July 10, 1905, to third unknown date	Retreat continued
Third unknown date to July 15, 1908	Advance
July 15, 1908, to June 24, 1909	Advance continued
June 24, 1909, to August 23, 1909	Advance continued
August 23, 1909, to July 4, 1910	Advance continued
July 4, 1910, to September 5, 1910	Advance continued

The Eastern Ice Cliff. The eastern ice cliff of Columbia Glacier extends from the morainic islet in the cove, described above, northeastward to the mainland. It is a mile and a quarter in length, very sinuous, and rises approximately two hundred feet above the water in a pinnacled and crevassed cliff whose white or blue clearness is not obscured by morainic load except for certain black basal layers and a lateral moraine strip on the east near where the glacier reaches the mainland. This eastern cliff changed very decidedly from 1909 to 1910. In the former year it was almost exactly as in 1899, 1905, and 1908; evidently rising out of shallow water and resting on bottom, for, although the cliff was precipitous, there is no great activity of iceberg discharge. All the bergs were small, and all were apparently discharged by sliding down the ice front. Photographs taken from Gilbert's 1497 foot station (I), on the hillside to the east show that this part of the eastern ice cliff advanced slightly between 1899 and 1909.

In 1910 the ice cliff had advanced 150 or 200 feet more, pushing up a terminal moraine in front of it so that except in the western half mile no part of the eastern ice cliff ended in tidewater. Even the western portion did not reach the bay proper but ended in a cove, for the tidal cliff is fronted by the low barren islet previously alluded to. Here the water is evidently deep and small icebergs are occasionally discharged. The water in front of the eastern half of the cliff is shallow, being only 57 feet deep at a distance

[1] Perhaps interrupted by a slight readvance, about 1897 or 1898, suggested by the small inner push moraine observed by Gilbert in 1899 on the western shore of the bay, on the islet north of Heather Island, and on the mainland to the east. The fluted ground moraine beside Heather Island terminus also suggests an advance between 1892 and 1899, as Gilbert says (Op. cit., p. 78).

of a half mile, and even shallower near the glacier, which has therefore been able to plough up the mud of the sea bottom into the new terminal moraine that rose above the bay in 1910. In this eastern half, therefore, no icebergs were discharged, as was the case in 1909, and a few streams flowed into the bay across the mud moraine. Only on the extreme eastern edge of this ice cliff was the new terminal moraine made up of anything but mud. There the beach gravels were being pushed up into a ridge 200 feet long, 130 feet wide and 20 feet high, and filled with mussel and other shells, with barnacles attached to pebbles, and with seaweed. In front of this push moraine were three folds of the compact clay, forming ridges parallel to the front of the moraine against which the ice was advancing. The one nearest the push moraine was breached axially at the crest.

Between July 2 and September 6, 1910, there was sufficient advance along this margin of the glacier to dam back a portion of the tide flats of early summer into a small lake.

The Lobate Eastern Margin. The eastern margin of Columbia Glacier is a sinuous lobate terminus ending on high and low land of the alternating rock hills and plains. Along this eastern margin Gilbert described phenomena in 1899 similar to those which he observed on Heather Island. "There was an inner push moraine, chiefly or wholly of drift and running parallel to the ice margin. There was an outer push moraine, less regular in its distance and associated with disturbance of the forest and the meadow peat. In the tract between the two many prostrate trunks were seen, showing that in places the front of the forest had been crowded back several hundred feet. Many of the trees that were overturned but not overridden, retained their bark, branches, and even minor twigs, but the leaves had fallen. On disturbed forest soil Coville found three young spruces which had grown since the catastrophe. In each case the age, as shown by rings of growth, was seven years. The date of the ice maximum was therefore not later than 1892 and may have been that year."

We do not know the behavior of this margin during the earlier oscillations described for the Heather Island and western termini. Advance had taken place here before June, 1909, however, for Grant states that "on the eastern edge of the glacier there was a zone of perhaps 200 feet between the maximum advance in recent years into the forest and the front of the ice." [1]

On August 24, 1909, the National Geographic Society's expedition was able to compare the exact conditions along the lobate margin with those ten years before. We climbed the spur east of the glacier to Gilbert's plane table and photograph station (Sta. I), at an elevation of 1497 feet; and we also walked some distance along the marginal stream and lake southeast of the glacier, but did not walk over any of the glacier edge itself.

On comparing the photographs from this spur taken by Gilbert on June 26, 1899,[2] with the conditions on August 24, 1909, we found the following changes. At the margin of the eastern ice cliff there was an advance southwestward into the bay which we estimate to be between 1000 and 1500 feet. There was a broadening of the glacier by several hundred feet, as also shown by Grant's observation in June, 1909, but in August the ice margin had as yet not extended to the maximum of 1892. In places, however, it was within a short distance of the forest.

[1] Personal communication.

[2] Reproduced as Plate IX, Harriman Alaska Expedition, Vol. III, 1904.

The lobate ice margin spreads out in the depressions between a series of rock hills along the eastern margin and rides up on the sides of the hills. The position of the 1892 advance was clearly marked along this whole margin but the ice did not extend quite up to the windrows of prostrate and inclined trees, though since 1899 it had advanced several hundred feet and overridden the sites of several small ponds.

During the last week in June and the first week in July, 1910, we observed the conditions along this lobate east margin in detail. There had been an advance of two or three hundred feet so that the ice was up to or just beyond the 1892 maximum, having completely covered the barren zone.

Just east of the terminal moraine that fronts the eastern ice cliff, which was tidal ten months before, was an area of low treeless morainic hills made up of glacial deposits, in places resting upon rock ledges. In 1910 the glacier was ploughing forward into this deposit and had developed a series of minor lobes. There was no pronounced terminal moraine at the ice edge but there was a low push moraine, two or three feet high in places. The glacier was evidently crowding the whole deposit forward against the rock slope to the southeast. Locally near the ice edge folds of peaty material had been rolled up, but throughout most of the area, there was no covering of peaty soil upon the glacial till and, therefore, no folding.

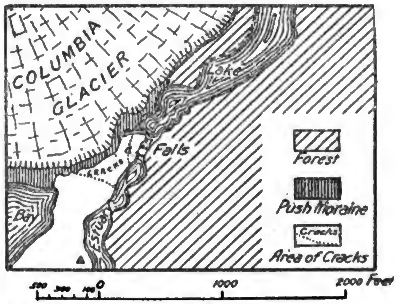


Fig. 34. Portion of Lobate Eastern Margin of Columbia Glacier, Showing New Lake Formed Between August, 1909, and June, 1910.

This is on site of "large glacial river" on Grant and Higgins map (Fig. 31) made in June, 1909. At *a* was an abandoned river course 24 feet above high tide in June, 1910, and marine fossils were found on top of hill. Dotted lines enclose area of cracks near waterfall where advance was pushing up a dam of glacial till in 1910.

The mass of till had been pushed forward against the rock slope to the south and had then bulged upward, damming back the stream into a new lake (Fig. 34), at whose outlet there was a foaming waterfall 10 feet high in June, 1910, where there was none in August, 1909. This lake rose high enough to submerge trees, shrubs, moss, and turf that in the fall of 1909 were growing high above the water. Other evidence of the bulging upward of this till mass was the presence of marine shells and seaweed on top of a hillock beside the waterfall, 24 feet above high tide. These shells were within the reach of the tide ten months before, for at high tide in 1909 salt water went up the river a quarter mile above the site of the waterfall of June, 1910. It was a narrow estuary at high tide and a narrow stream at low water.

The surface of this whole disturbed area was gashed by gaping cracks, most of which were shallow, due to the compression and upward swelling of the till mass under the thrust from the advancing glacier. A majority of the cracks extended at right angles to the front of the glacier, that is parallel to the direction from which the strain was being

applied. Many of them were merely due to the cracking of the thin, grassy cover of the till, and across quite a number still extended fine threads of grass roots which had been drawn out as the turf on either side was stretched and broken. The whole arrangement was characteristic of an expanding mass whose turf cover fitted it before the strain was applied but which had since bulged upward so that the turf covering was too small.

Over the top of the 24-foot hillock which dammed back the lake ran the abandoned channel (a, Fig. 34) of the marginal stream of 1909, its site marked by rounded stream gravels and by the absence of vegetation. In late June and early July, 1910, the stream course at the outlet of the lake was at the outer edge of the till mass where it was being pushed up against the mountain side.

It was also evident that there had been several minor lake stages preceding the current one, for on the lakeward slopes of the 24-foot hillock were minute, abandoned shorelines. They were not horizontal, however, but inclined.

The continuation of this crowding forward of the till dam resulted in several changes in drainage and topography along this part of the glacier margin between July 2 and September 6, 1910. Several of the minor lobes advanced 20 to 40 feet and there was much more fracturing of the turf than in July, the turf was being pushed out into the river just below the falls where there was only a mud shore in July, the course of the lake outlet was changed, the ten-foot waterfall of early summer being abandoned, the lake level had been lowered slightly by the downcutting of the new outlet which had two channels instead of one. These channels had cut away a large part of the 24-foot hillock, destroying portions of the abandoned 1909 stream course, and the new waterfalls were only 5 or 6 feet high.

The site of Gilbert's 1899 photograph of the glacier margin described on p. 276 was immediately north of the marginal lake just described. In June, 1910, the ice had advanced clear to the 1892 maximum, several of the prostrate trees killed at the earlier date being still recognizable; but by September they had been overridden. Along this portion of the glacier margin the windrows of dead trees killed in 1892 were still undisturbed in August, 1909. They lay in regular piles, generally with their tops pointing away from the glacier. In June, 1910, the ice had advanced up to many of them and they were heaped up in irregular masses, forming a tangle of wood, turf, and rock débris in one place 75 or 100 feet high.

The changes along this margin of the glacier will not be described in detail, although shown in several photographs. It is enough to say that there were great changes from August, 1909, to June, 1910, and from then to September, 1910. The changes varied, depending on whether the ice was ploughing up turf or forest, and whether it was advancing up a hillslope or down a declivity. All of the barren zone which Gilbert described in 1899, was overridden in 1910, so that forest not touched in 1892 was then being destroyed all along the margin, excepting possibly on a terminal moraine described later.

There were two lower lakes along the eastern margin of Columbia Glacier. In 1909 the lower or westernmost and smaller of these extended eastward from a low waterfall over a rock ledge a quarter of a mile east of the till dam pushed up in 1910. Below this there was no lake in 1909, and as already stated, tidewater then extended up to these falls, three quarters of a mile from Columbia Bay. The till dam of 1910 made a new lake, both longer and deeper than the older one, covering the site of it and of the low water-

fall alluded to. In 1910 the edge of the advancing glacier extended into this new lake in two lobes a short distance apart.

At the western end of the upper and larger lake, which remained unchanged in 1899, 1909, and 1910, there is a higher waterfall over rock. In this upper lake the glacier margin turns from its northeast trend and continues northwestward. Here the 1892 maximum was marked by a low crescentic moraine from which the glacier had retreated so far in 1899 that there was a line of crescentic pools between the moraine and the ice front. The readvance at this point in 1909 had extended out across the site of these pools and nearly out to the 1892 terminus. We estimated the advance in this lobe to be nearly a quarter of a mile. Between August 24, 1909, and June 30, 1910, the continued advance of this lobe carried it out about 300 feet and there was a lateral spreading eastward of 330 feet. The 1892 moraine was completely overridden in places, so that the ice extended into the lake in three minor lobes where none reached the water ten months before. This was also true in another minor lobe to the southwest. An interesting little delta, evidently built in this lake in 1892 when the crescentic moraine was made, shows in Gilbert's 1899 photograph as a small abandoned delta with no streams except the small outlet of a marginal lakelet. In August, 1909, the ice extended nearly up to the delta, though not far enough to send streams across it; but in June, 1910, the advance had brought the ice up to it, and outwash streams were once more building it forward into the large lake.

Along the more distant lobes, that project against the forested rocky knobs still farther northward, holding marginal lakelets there, the ice seemed in 1909 to extend right up to the forest and had apparently quite equalled in extent the earlier great advance in 1892. In 1910 there had been additional spreading of the glacier so that several of these lobes had advanced 200 to 400 feet into the forest.

Along this lobate border there were many modifications in 1909 and 1910. The glacier had thickened, crevassing had sliced up ice that was smooth in 1899, and the distribution of moraine upon the margin had changed somewhat from 1899 to 1909. Although there was more crevassing in 1910 than in the previous year the distribution of marginal ablation moraine was about the same.

The great lobe, or distributary, east of the nunatak, underwent no appreciable change from 1899 to 1910 and we assume that it still coalesces with the main glacier below the nunatak, as Palache observed in 1899. We saw this distributary from too great a distance to be sure as to the detailed conditions in 1909 and 1910, but, as the topographic map shows, it touches the main glacier but contributes no ice to it.

The drainage of the eastern margin of Columbia Glacier shows an interesting alternation of complex conditions. The streams emerging from the eastern glacier distributary, joined by the drainage from the eastern margin of the main glacier south of the big nunatak, and by smaller streams from small mountain glaciers on the east, flowed southward into a lake among the rocky hills east of the main glacier. This lake was nearly filled by a delta of outwash gravels which grew forward appreciably between 1899 and 1909. From this upper lake the marginal stream flowed southward through the rocky, forested hills east of the glacier into the long, crooked, upper lake. The main mountain wall formed one shore of this lake, and the moraine bordering the minor glacier lobe that advanced a quarter mile between 1899 and 1909, formed the other. Into this muddy lake flowed numerous turbid streams from the glacier, some of which built deltas.

Below this lake the stream course was partly marginal to the present ice sheet for a short distance, then came several sections in broad gorges through the drift and rock with some waterfalls, alternating with broader, lake-like sections where only the outer moraine separated the stream from the present ice edge. Here the lobate eastern margin of the glacier formerly forced the stream into marginal channels across rock spurs, making lakes above the spurs and rapids on the rocks, and these conditions still lingered, though the stream had cut somewhat in the rock. Below the westernmost gorge the stream emerged into a muddy tidal estuary which branched from the bay east of Heather Island in 1909, while across a part of it in 1910, the advance of the glacier had formed another lake with a till dam, as already described.

The Main Glacier Surface. Columbia Glacier surface slopes about 150 feet to the mile. The glacier margins discussed in the preceding pages represent only the outermost portions, and do not include those in the upper part of the glacier at a distance back from the sea. These portions of the glacier, and the low grade, main glacier surface, we saw only from the crest of Heather Island and from the 1497 foot station southeast of the glacier. The conditions can be described fairly well, however, for the viewpoints were comprehensive and we were able to write notes and take photographs from stations at which similar records had been made ten years before.

The main glacier surface was impassably crevassed in each of the periods of observation, though it was more severely crevassed in 1909 and 1910 than ten years before, and on the east side there were one or two ice steps with especially severe crevassing. The changes in the medial and lateral moraines, and in crevassing within them, are the most important differences between the 1899 glacier surface and that which we saw, for the changes accompanying the thickening and spreading of the margins were not determinable in the main glacier back of the frontal portion.

Gilbert has described the conditions in 1899 as follows:[1]

"Opposite the great nunatak were two medial moraines, one passing within a half mile of its base, the other lying about one mile from the opposite edge of the glacier. A central tract two miles broad was practically drift-free. Toward the end of the glacier this central tract broadened, the medials swinging toward the sides, until finally the white belt was three miles wide. As the medials diverged they also broadened, and they eventually merged with flanking moraines, so that near the end, especially on the east side, the areas of drift-covered ice were very wide. The marginal belt on the west, instead of continuing northward parallel to the medial with which it was associated, was seen to curve about into the western embayment, as indicated on the map, and a belt seen from a distance near the north edge of the embayment was supposed to be its continuation, a loop being made within the embayment. As the ice in the embayment descended toward the west, it is evident that the morainic loop could not at that time represent a line of continuous flow, for we cannot suppose the ice to flow into the embayment on a descending course and then return on a parallel ascending course. It is therefore probable that the moraine was formed as a comparatively direct line of drift, following the course of the main current at a time when no current entered the embayment. The inference that a change has occurred naturally leads to inquiry as to the precise nature of the antecedent condition of the glacier. On the one hand, the embayment may have been so full of ice that the surface gradient was outward; or, on the other, the glacier of the main valley

[1] Op. cit., pp. 74–5.

may have had so low a surface that there was no tendency to overflow to the comparatively shallow side valley.

"The first case implies snow accumulation in the embayment a few decades ago at a rate not since maintained, and would correspond to a general expansion of glaciers followed in later decades by contraction; but the relations of the ice to the forest, to be described presently, show that such contraction has not taken place. The second case implies a general expansion of the glacier as the important element of its later history.

"Another medial moraine of the great ice field north of the nunatak passed just east of the nunatak and continued down the eastern arm of the glacier to its end, where it contributed toward the building of a great alluvial delta which was gradually obliterating one of the lakes."

In 1909 the band of clear ice between the east medial and the lateral moraine was narrower than ten years before, as if advance in midstream had crowded it over. The same thing seemed to be true of the whole broad east moraine below the junction of the medial and upper lateral moraines. Where it had been smooth in 1899, this broad lateral moraine was in 1909 severely crevassed throughout, and some débris had fallen into the crevasses; but the contact of clear and crevassed ice was as sharp as ever and the lateral moraine belt was narrower. We, therefore, infer lateral spreading of the area of clear ice in the interim. In 1910 this lateral spreading and crevassing had increased slightly, but there was only a little more white ice on this east side than the year before.

The moraine on the distributary east of the large nunatak, was slightly more extensive in 1909 than ten years before; but since snow covered it in June, 1910, we cannot tell its condition then. The apparent change from 1899 to 1909 may be due in part to more snow on this upper part of the glacier during Gilbert's June observation than at the time of our observations in August. This ice tongue is much less crevassed than the main glacier, although it has a steeper grade. It has scattered débris upon its eastern and terminal margins, less than in 1899, but the western side has a broad lateral moraine, apparently heavier than ten years ago, which seems to be made up partly of marginal débris from the large nunatak, but chiefly of material carried by a medial moraine from the main upper glacier. Near the northern end of the nunatak this medial moraine joins the lateral moraine, which had broadened and extended farther out into the ice in 1909 than ten years before. This extension is perhaps due to ablation near the warmer land.

On the western margin of the glacier we were unable to look far enough into the southermost embayment to see whether the moraine still bends into it, as observed by Gilbert and shown upon his map. There seemed to be a marginal lake with icebergs on the northern side of this embayment in August, 1909. On June 30, 1910, this lake had been drained, its site being marked by heaps of tumbled icebergs which had gone aground when the water was drained out beneath the main glacier.

In both this embayment and the one next north, the grade is so flat that it could not be determined from our remote viewpoint, whether the grade was reversed, as inferred by Gilbert, or whether these are debouchures of tributaries.

Still farther to the north and northeast the main glacier receives a number of accordant, hanging, and cascading glacier tributaries; but the absence of pronounced medial moraines away from the margins of the glacier indicates that though many of these are good-sized glaciers, they are all relatively small and unimportant compared with the main ice tongue.

Glacial Erosion. A sounding a little over a quarter mile from the terminus of Columbia Glacier showed 600 feet of water, indicating clearly that the ice front is not afloat. It is, therefore, grinding actively upon the bottom of the fiord. Other soundings (Fig. 35) show clearly that it has not been afloat at any time when it extended farther south in Columbia Bay, the depths of water increasing gradually from 600 feet near the ice front

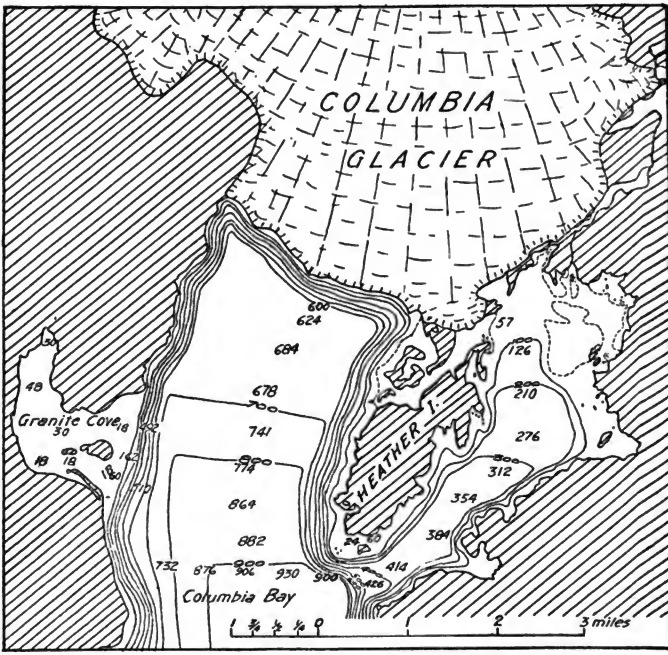


FIG. 35. SUBMARINE CONTOURS IN FRONT OF COLUMBIA GLACIER.

to 906 feet at a distance of 3½ miles and 1020 feet five miles from the ice front at the entrance to Prince William Sound, the slope being uniformly about 250 feet to the mile. As the soundings, a half mile apart, show no interruptions in this even grade it is evident that there has been no significant basining of the fiord bottom. This would be normal in a straight section of fiord without constrictions of the side walls. In the narrower channel east of Heather Island the water is 57 feet deep a half mile south of the east ice

cliff and the water also deepens uniformly southward but at the rate of only 98 feet to the mile, the depth 3¾ miles from the glacier being 426 feet (Fig. 35).

The cross section of the fiord (Fig. 36) is of a very broad U-shape, with broad flat bottom, typical of glacially-sculptured troughs.

There are submerged hanging valleys on each side of the main fiord, the channel east of Heather Island hanging 474 feet above the bottom of Columbia Bay, while the smaller Granite Cove, at the entrance of which the water is only 18 feet deep, hangs 756 feet. There are also hanging valleys above sea level, one about half way between Granite Cove and Columbia Glacier, on the west side of the fiord, hanging 300 feet above the level of the water, which is here nearly 700 feet deep in midchannel. The whole western slope of the fiord is much over-steepened by glacial erosion, having a slope in many places of over 4000 feet to the mile, with sheer precipices in places and with much bare rock. Cirques, hanging valleys, roches moutonnées forms, over-steepened lower slopes, and truncated spurs are also abundant in the mountains on the eastern side of the fiord, but this side has been less sharply eroded than the western. Glacial striae are rarely seen in Columbia Bay, for soil and moss mask most of the rock ledges.

Glacial Deposits. Glacial deposits form no conspicuous features near Columbia Glacier, and there are no submerged deposits suggestive of moraines in the bay. Heather Island

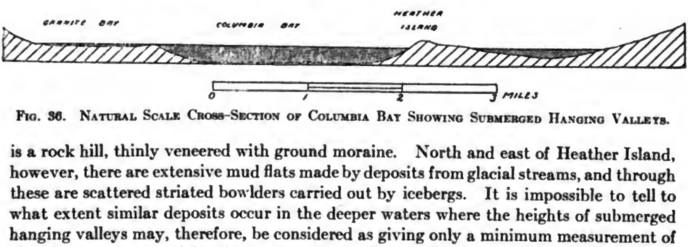


FIG. 36. NATURAL SCALE CROSS-SECTION OF COLUMBIA BAY SHOWING SUBMERGED HANGING VALLEYS.

is a rock hill, thinly veneered with ground moraine. North and east of Heather Island, however, there are extensive mud flats made by deposits from glacial streams, and through these are scattered striated bowlders carried out by icebergs. It is impossible to tell to what extent similar deposits occur in the deeper waters where the heights of submerged hanging valleys may, therefore, be considered as giving only a minimum measurement of glacial erosion. The lakes in the valleys southeast of Columbia Glacier may be due either to glacial obstructions or to basining by sculpture of tributary ice tongues. In the marginal lakes immediately adjacent to the ice there are some glacial deposits, among which the outwash plain of sand and gravel built by the stream from the large glacier distributary on the east is conspicuous. It is over a mile in length.

The lowland next the bay, over which the lobate eastern margin of the glacier was beginning to advance in 1910, is the only extensive area of terminal moraine in the region. It is five-eighths of a mile long, an eighth of a mile or less in width, and has the sharp undulating knob-and-basin topography (Pl. CXII, A) typical of many terminal moraine areas in the United States. There are a score or more of small ponds in the kettles. Most of the area is covered only by grass, but the parts farthest from the glacier have very small strips of mature conifers with dead trunks heaped up along a push moraine at the border of the westernmost clump. As Grant and Higgins have pointed out [1] the shrubs on the grassy moraine surface are not over 20 years of age, or thereabouts, but we do not agree

[1] Grant, U. S. and Higgins, D. F., Glaciers of the Northern Part of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLII, 1910, p. 729 and Fig. 8.

with them in believing that vegetation does not gain a foothold almost at once on areas uncovered by glacial ice where unconsolidated deposits are present and seeding possible. They have estimated that this advance took place 50 years or so ago, while we are more inclined to ascribe the building, or at least the last modification of this moraine, and the destruction of these trees, to the advance that occurred about 1892 or to an earlier advance only a few years before that date, and certainly not as long ago as 1859.

A section revealed in this glacial deposit by a wave-cut cliff near the western end shows the following succession, indicating at least two advances here.

SECTION OF GLACIAL DRIFT NEAR COLUMBIA GLACIER

Present surface, with morainic topography and growing shrubs.
Till, with blue clay and angular, striated stones.
Trunks of trees, and cones and fragments of bark.
Till, with large angular glaciated bowlders.
Bedrock, with glacial striae.

The lower till surface was an undulating one, evidently produced by irregular erosion of the lower till mass.

Former Stages of Columbia Glacier. There are many facts to prove that Columbia Glacier is now far less extensive than formerly, and that at an earlier stage it was a tributary of the great Prince William Sound glacier. Near Columbia Glacier the cirques, hanging valleys, roches moutonnées hills, over-steepened lower slopes, truncated spurs, the generally rounded topography produced by glacial sculpture, and the glacial deposits testify to this former expanded stage of glaciation. How thick the ice was at maximum has not been determined exactly. Heather Island, 357 feet high, was glaciated to the top. The mountain spur southeast of the ice tongue was glaciated as high as we climbed it, the elevation of the photographic station at timber line there being 1497 feet. From a distance the great nunatak has the rounded appearance of having been glaciated clear to the top, 3664 feet, though this is not certain. Gilbert places the height of glaciation in Columbia Bay at 4000 feet.[1]

Between this great maximum and the present stage of expansion, and with unknown intermediate episodes, there has evidently been notable recession of the glacier, proved by forests upon the mountain spurs above the upper glacier for a distance of at least six or seven miles from the present terminus, and upon the lower slopes of the great nunatak. It is conceivable, though hardly probable, that the forest may have migrated to the nunatak along the eastern ice margin with the glacier as expanded as now; but it is much more difficult to believe that the forest on the western side has migrated so far up the margin of a clear ice glacier. The farthest trees seen were on the northern side of the second embayment, or tributary, on the western side of the glacier. They seemed to be spruces. The greenish tint of the lower slopes beyond the next tributary suggested vegetation there, perhaps shrubs like the alder, but no trees could be detected with the field glasses. This last locality is over ten miles from the present ice front. Gilbert interpreted the bending of moraines into the first embayment as evidence that the 1892 maximum was preceded by an important minimum,[2] and we are inclined to interpret this distribution of forest in the same way. Grant and Higgins have inferred a retreat

[1] Op. cit., p. 174.
[2] Op. cit., p. 80.

previous to 1892, though not necessarily of more than a mile, on the basis of the marine shells close to the ice front, assuming that these sea forms would not thrive in such cold, milky water.[1]

That the glacier has not recently been as far out as in 1892 and 1910, is proved by the extension of mature forest with trees at least 75 to 100 years old, up to the very border of the ice. Peat three or four feet thick has accumulated in grassy glades on hilltops and flattish ground, while living trees grow only on rock or other drained elevations and along stream courses, dead trunks standing where overwhelmed by the thickening tundra masses. These relationships of vegetation, which require time for their establishment, are found in Prince William Sound but not in the Yakutat Bay region, where the glaciers were recently much more expanded. Into these vegetation areas the advancing ice of 1892 and 1909–10 plowed its way.

Perhaps, as Gilbert suggests, the important minimum of Columbia Glacier that preceded the 1892 advance occurred in the nineteenth century, when Petroff does not mention a tidal glacier in Columbia Bay, though he does assign one to Port Valdez. As we have attempted to show in our preceding discussion of the subject, there are grave reasons for doubting the existence of a large tidal glacier in Port Valdez fiord at the time Petroff mentions.

Within a year or so of 1892 Columbia Glacier had an important advance, since which there have been two periods of retreat and two of advance, the last of which was still in progress in 1911.

The cause for these recent oscillations is not known. They may be climatic or they may be due to earthquake avalanching. If the latter is the cause, we should be inclined to ascribe the advances to avalanches during earthquakes in the Chugach Mountains, not to the Yakutat Bay earthquakes of September, 1899. Both climatic and seismic data are too meagre for settling this question, but in 1907–08 there was a 148 inch increase of snowfall[2] at Valdez, 25 miles to the east, (from 299 inches in 1906–07 to 447 inches in 1907–08). The 1906–07 snowfall at Valdez seems to be below the normal amount. It decreased to 187 inches in 1904–05. There was a 337 inch increase in snowfall at Valdez (Fort Liscum) during the winter of 1901–02 (see table at end of Chapter XV). On the other hand, there were severe earthquakes in this part of the Chugach Mountains in May, 1896, August, 1898, October, 1900, March, 1903, and February, 1908.

Since the 1892 advance, whose cause we will not discuss, the first of the recent advances came between 1899 and 1905, and may have been due to the 1896 or 1898 or 1900 earthquakes. The advance which was in progress at the time of our visits in 1909 and 1910 probably began in 1907 and 1908 and may have been caused by one of the preceding earthquakes, or, as already suggested, it may be that both advances were caused by variations of snowfall or of temperature. In the latter case other adjacent glaciers should have advanced and retreated, for notable temperature variation can hardly be as localized as precipitation may be.

In view of the limited data at hand it is not profitable further to discuss the cause of the fluctuations of Columbia Glacier, though we are convinced that it is well worth while

[1] Grant, U. S. and Higgins, D. F., Glaciers of the Northern Part of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLII, 1910, pp. 729–731.

[2] Martin, Lawrence, Zeitschrift für Gletscherkunde, Band VII, 1913, pp. 28-31.

to call attention to the possibility of explaining them by earthquakes. There is, however, one very notable difference between the oscillations of Columbia Glacier, and the advance of the Yukatat Bay glaciers in response to the 1899 earthquakes. The latter advanced spasmodically and rapidly, they were transformed in a few months' time, and the advances were short lived; the Columbia Glacier has moved more slowly, continuing its advance a longer time and gradually increasing its rate of forward movement from nine-tenths of a foot a day in 1908 to $2\frac{1}{10}$ feet a day in 1910. This less abrupt change is more like that of the glaciers in regions like the Alps, where climatic fluctuation is assigned to explain the advances and retreats. Even this difference, however, need not eliminate the earthquake theory, for the rate and extent of the advance in response to earthquake shaking would normally vary with the intensity and duration of the shaking, as it would with variation in the amount and duration of a climatic change. Such spectacular response as has been shown by the Yakutat Bay glaciers is hardly to be commonly expected. We believe the earthquake theory a possible and rational explanation of the recent fluctuations of Columbia Glacier front; but, as between it and the climatic theory, we have no facts upon which to either eliminate the one or establish the other.

CHAPTER XV

THE GLACIERS OF UNAKWIK INLET AND COLLEGE FIORD

Unakwik Inlet

General Topography. Unakwik Inlet is a large fiord west of Columbia Bay. It is about 20 miles long, trending north and south, and having an average width of about 2 miles. On the west side of it are Jonah Bay and Siwash Bay, and on the east is a large cove, at the head of which is Miner's River. About 2 miles from the head the inlet narrows to a half mile and turns at right angles nearly eastward. Here it is terminated by a large ice tongue, Meares Glacier, the only glacier in this fiord which reaches tidewater; but Brilliant Glacier, Ranney Glacier, and a number of smaller ice tongues are visible from the inlet and send streams down nearly to it.

Meares Glacier. This ice tongue (Fig. 37), which has a known length of about 6 miles, is made up of two smaller glaciers (Pl. CXII, B) which unite a mile and three-quarters from the terminus. It is a clean white ice stream, severely crevassed throughout its visible extent, and carrying one rather-weak medial moraine, which comes from the junction of the two chief tributaries, and is pushed over so near the south side of the glacier that the north tributary is shown to be larger and more vigorous than that from the east. There is a strong lateral moraine on the north side but none on the south. At tidewater the glacier has a width of one mile, and a terminal cliff over 200 feet high, from which so many icebergs are discharged that, at the narrows in Unakwik Inlet, it is usually difficult to force a boat through them. There are several very pretty cascading glaciers on the mountain slopes near Meares Glacier.

The first known description of Meares Glacier was made in 1790 by the Spanish explorer Fidalgo, who did not name the glacier, but on the chart of the *Sutil y Mexicana* called Unakwik Inlet by the name Puerto de Revilla Gigedo. Don Francisco Eliza, the commander of this expedition, refers to the glacier as "Volcan de Fidalgo." Fidalgo's description of the region as he saw it on June 15, 1790, quoted by Davidson from a manuscript, is as follows:[1]

"So soon as they arrived at the mouth of a sheltered harbor, almost at the northernmost part of the bay, they observed the latitude to be 60° 54′, and heard many subterranean explosions or thunderings which increased as the sun approached the meridian. Conducted by their Indian guides to the interior of this harbor, they discovered in its depths to the north, a great level tract of snow which came to the water's edge, and ended at the base of the high mountains. Hardly had they seen this, when they noticed that with each subterranean roar a mass of snow was thrown up from the center of the plain about half the size of the launch; fearful lest they should be

[1] Davidson, George, The Glaciers of Alaska That are Shown on Russian Charts or Mentioned in Older Narratives. Trans. and Proc. Geog. Soc., Pacific, Vol 3, 1904, pp. 32–33.

overwhelmed or destroyed in this port, they did not continue their examination of that phenomenon, which is undoubtedly worthy the attention and investigation of a naturalist."

Unakwik Inlet was visited by Lieutenant Whidbey of Vancouver's expedition on June 9, 1794, the description being as follows: [1]

"This arm was found to take a north direction, in general about a league wide, and to terminate at the distance of about four leagues, at the foot of a continuation of the range of lofty mountains before mentioned. Its upper parts were much incumbered with ice, as were both the eastern and western sides with innumerable rocks, and some

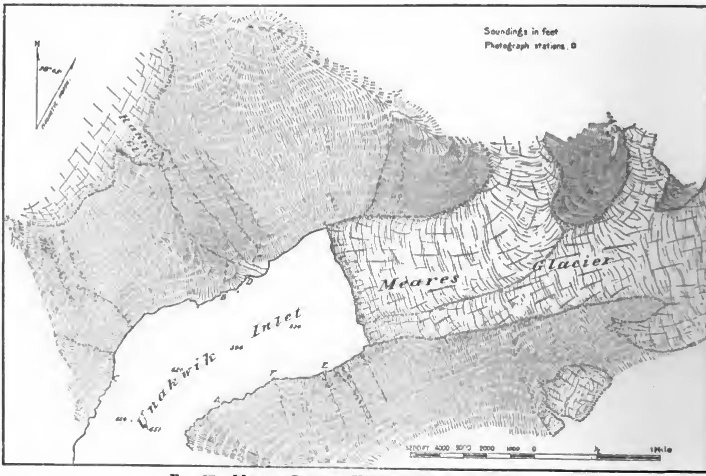


FIG. 37. MEARES GLACIER, UNAKWIK INLET, IN 1910.

islets." The mention of ice in the upper part of this bay shows that Meares Glacier still reached tidewater in 1794, though Whidbey probably did not go far enough to see it.

In 1905 and 1909, U. S. Grant went to the head of this fiord, and in the latter year Grant and Higgins mapped the inlet and glacier, which they named.[2] They show that Meares Glacier is nearly twice as long as was indicated on previous maps,[3] heading

[1] Vancouver, George, A Voyage of Discovery to the North Pacific Ocean and Around the World, Vol. V, London, 1801, p. 316.

[2] Grant, U. S. and Higgins, D. F., Glaciers of the Northern Part of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLII, 1910, pp. 736–738; Journ. Geol., Vol. XI, 1906, p. 407; Bulls. 443, U. S. Geol. Survey, Pl. II, and 379, Pl. IV.

[3] U. S. Coast and Geod. Survey, Chart 8550, 1909.

nearly as far north as Yale Glacier instead of at a point nearly east of Crescent Glacier in College Fiord.

The National Geographic Society's expedition of 1910 spent July 12th to 14th in the study of Meares Glacier and Unakwik Inlet, also making the hachure map of the glacier reproduced as Fig. 37 and the soundings shown on Fig. 38.[1]

Grant has stated [2] that in 1905 "the bushes and trees were close to the ice and there was no bare zone, or at most a very narrow one, visible between the ice and the forest. In 1909 the front of the ice seemed to be a little in advance of its position of four years before. At the later date near the front of the glacier on the south side was a brown zone estimated to be 200 feet in width. This brown zone appeared to have been caused by dead vegetation rather than by bare rock, and at the edge of the ice there were a few small trees. Close to the glacier there was a sparse forest which contained trees estimated to be ten inches in diameter. Hence the ice was probably as far forward in 1909 as it has been during the last hundred years."

In 1910 the north edge of the glacier (Pl. CXIII) had advanced slightly since the year before, as was shown by a comparison of photographs and by the nearly vertical terminal ice cliff on the beach, at the base of which were freshly-fallen ice blocks. Near the southern side we were unable to determine positively whether the glacier had remained stationary from 1909 to 1910, or whether there had been a slight retreat. A mass of ice, clinging to the mountain side at sea level, was connected with the glacier in 1909, but by July, 1910, had been detached by melting. It is easy to understand the advance of one side of this glacier and the stationary condition or slight retreat of the other, in view of the fact that the tributary from the east is relatively inactive, while the one from the north was more active in 1910 than in the year before.

This difference in activity of the tributaries is shown by the conditions along the margins of the glacier below their junction. On the south margin there was no change in condition during the last two years of observation, while along the north margin the glacier was spreading as well as advancing, everywhere extending up to the forest and uprooting the turf, in places pushing back into and even overriding the forest. These conditions were observed both at the terminus and up to the turn in the glacier. Slivers of ice projected among the alders and spruces and small push moraines of till and turf were seen. Several spruces, not yet overturned, stood at the very ice margin, here stained dark colored with lateral moraine.

In the quotation given above, Grant refers to a brown zone of dead vegetation on the south margin of the glacier in 1909, and this was not overridden in 1910. It may be interpreted as the result of forest destruction during an advance of the east tributary before 1909. One mature tree, tilted during the advance, stood at the edge of this brown zone and there were several young shrubs at the very edge of the ice.

The terminus of the glacier was precipitous in 1910, as if the end were in deep water, and at a distance of a half mile the fiord is 534 feet deep, suggesting that the end of the glacier is not afloat.

We agree with Grant that Meares Glacier is probably as far advanced now as it has been in the last century. There is considerable forest on the mountain slope above the

[1] Martin, Lawrence, The National Geographic Society Researches in Alaska, Nat. Geog. Mag., Vol. 22, 1911, p. 551.

[2] Bull. Amer. Geog. Soc., Vol. XLII, 1910, p. 738.

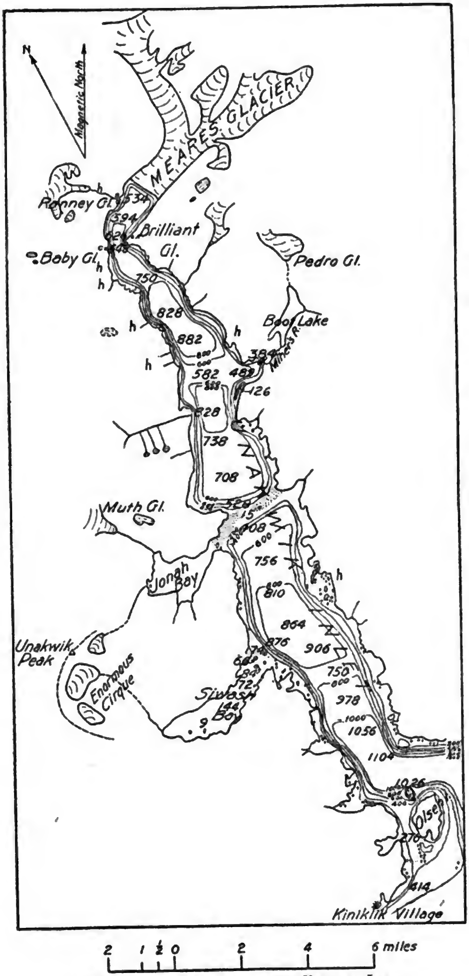


FIG. 38. SUBMARINE CONTOURS IN UNAKWIK INLET. Outline of coast and glaciers after Grant and Higgins.

glacier, and on its north side, but on the south side the thick forest ends almost exactly on a line with the edge of the ice, there being only a very few scattered trees beyond. There is some alder on the spurs between the two tributaries, but no spruce.

Ranney, Brilliant, Pedro, and Smaller Glaciers. Ranney Glacier of Unakwik Inlet is a cascading ice tongue (Fig. 37) west of Meares Glacier. It ends a thousand feet or more above the level of the fiord, as do Baby Glacier and several small unnamed ice tongues.

Brilliant Glacier is a cascading glacier on the east side of the fiord (Fig. 38), of nearly the same size and appearance as the well-known Illecillewaet Glacier in the Canadian Rockies. Instead of ending high above the fiord, however, as the Ranney Glacier does, it extends down a steep slope nearly to sea level. From its terminus a stony outwash plain and delta extend southwestward nearly a mile and a half to the fiord. In this valley bottom and on the delta are mature spruces.

Nothing is known about previous advances or retreats of these small glaciers, though the fact

that spruce extends up to the lips of several hanging valleys on the west side of the fiord shows that ice tongues have not extended out of these valleys for a century or more. Absence of spruce on spurs between hanging valleys may be attributed to snowslides. Ranney Glacier has evidently been retreating for several years because there is a large barren zone in front of it, marking the distance to which the glacier had advanced during the last maximum. Mature alders extend up to this barren zone.

Pedro Glacier heads in snowfields south of Brilliant Glacier terminating 3½ miles east of Unakwik Inlet and sending a small stream through two lakes to the cove of Miner's River. Muth Glacier, and three small ice tongues in an enormous cirque at the base of Unakwik Peak, terminate 1½ to 2½ miles west of the heads of Jonah and Siwash Bays respectively.

Glacial Modifications of the Fiord. Throughout Unakwik Inlet there is evidence of profound glacial erosion at a time when the Meares Glacier was much larger, receiving Ranney, Brilliant, Pedro and many small ice tongues as tributaries, and occupying all of Unakwik Inlet, forming one of the great tributaries of the piedmont glacier of Prince William Sound. There is also evidence of local erosion by the former tributaries of the Unakwik Inlet Glacier; for example, in the enormous cirque east of Unakwik Peak which has been cut back so far as to form a narrow arrête between Unakwik Inlet and College Fiord on which Unakwik Peak is the highest point. A similar narrow arrête has been developed between Meares and Brilliant Glaciers.

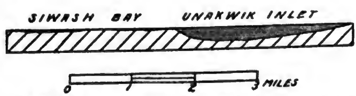


FIG. 39. NATURAL SCALE CROSS-SECTION OF UNAKWIK INLET.

Submerged hanging valley of Siwash Bay.

The main channel of Unakwik Inlet has the typical fiord topography, that is, it is straight and simple and has no spurs entering from either side, as a stream-eroded valley does. Instead, the spurs are all cut off with triangular facets and the fiord walls are eroded smoothly, although the width of the fiord varies from place to place. At the point where it is narrowest, in the turn west of Meares Glacier, one wall of the fiord is very much steeper than the other. As might be expected, the west wall is the steeper, this being the side on which the ice must have eroded most efficiently while the glacier was making this abrupt turn. The spur east of this curve slopes more gradually.

The tributary valleys that enter the inlet bear various relationships to the present water level in the fiord. Some of them enter at or just below sea level,—for example, Miner's River. Others end in the air and have streams which cascade down from the lips of hanging valleys (h, Fig. 38). Ranney Glacier is in such a hanging valley, and there are several others south of it on the west side of the fiord, some of them with a discordance of several hundred feet. There are also two hanging valleys on the east of the fiord. Still other tributary valleys are occupied by bays and it was not known that these were hanging valleys until soundings were made in 1910. These soundings show conclusively that the hanging relationship exists in Siwash Bay (Fig. 39), which hangs 790 feet above the bottom of the Unakwik Inlet. Although soundings were not made in Jonah Bay, it presumably has the same relationship. The channel behind the long island just south of Miner's River, with a depth of 126 feet, and the channel between the mainland and Olsen Island, at the south end of Unakwik

Inlet, with a depth of 276 feet (Fig. 38), show differential glacial erosion in these narrow passages, for the main channel is deeper in each case. These small channels are analogous to hanging valleys, but are discordant at both ends.

In general, the soundings reveal Unakwik Inlet as a U-shaped glacial trough (Fig. 39) with about the same valley slopes below sea level as above. The fiord bottom, with depths of 534 to 1266 feet has a basined character, which is interpreted as, in part at least, due to glacial deposits.

The most striking of these irregularities is a barely submerged reef (Fig. 38) 10½ miles south of Meares Glacier and just north of the mouth of Jonah Bay. At this point (M, Fig. 40), a reef rises 700 to 750 feet above the fiord bottom and extends continously across the fiord, its presence being manifested at the surface by a pair of bars, each extending nearly half way across the inlet. Just west of the middle of the fiord, the channel has a maximum depth of about 15 feet at low tide, while at high tide the pair of projecting points is crossed by several channels. The presence of a shoal here was observed by Glenn in 1898 when his small steamer struck the reef in mid-channel.[1]

This shoal is thought to be a terminal moraine rather than a rock reef or a pair of sand spits built out from either shore, both because it rises out of deep water at a point where glacial erosion must have originally made the channel nearly if not quite as deep as it is to the north and south, and because the surface of the reef and the projecting

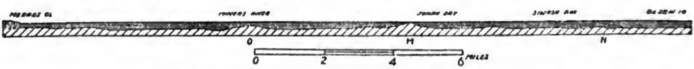


FIG. 40. NATURAL SCALE PROFILE OF UNAKWIK INLET.

points on each side of the fiord, where not covered by sand, are made up of rocks of variable sizes and shapes, bearing glacial striations, and including masses far too large to have been transported here by any agency except a glacier. If the shoal is made up of morainic material from top to bottom, the presence of a moraine over 700 feet in height shows that the ice front must have stood at this point for a long time in order to accumulate so large a deposit, for the accumulation of glacial material in the water is thought to be very slow. At the surface the bar differs from the terminal moraine on the land in the absence of unassorted clay; but even though made in the main of till the absence of clay in the upper portion is easily explained by the tidal currents which Glenn estimated to run at the rate of 10 knots an hour. These might easily have washed away the clay and left only the large rocks, upon which sand spits have been built from either shore. On the basis of this interpretation the deposit may be referred to as a *moraine bar*.

Five miles south of this moraine, and just south of the mouth of Siwash Bay is a second rise in the bottom of the fiord (N, Fig. 40). It rises 156 to 228 feet above the fiord floor, although the water is 750 feet deep on its crest. It is impossible to state definitely whether this is a recessional moraine or whether it is a reef of more resistant rock which was left by glacial erosion as a result of glacial scooping just to the north. If the latter, the depression to the north (756 to 906 feet deep) is a rock basin produced

[1] Glenn, E. F., Reports of Explorations in the Territory of Alaska, No. XXV, War Dept., Adj.-Gen. Office, 1899, p. 25.

by glacial erosion. If the low ridge, or swell, is a moraine it is an accumulation interrupting the normal slope of the fiord bottom which descends from a depth of 756 feet south of the Jonah Bay moraine, to 1104 feet at the entrance to Unakwik Inlet.

Four and a half miles north of the Jonah Bay moraine bar, and just west of the mouth of Miner's River, is a third submarine irregularity (O, Fig. 40), interrupting the slope of the fiord bottom, which is 534 feet deep at the end of Meares Glacier and 708 feet near Jonah Bay, with intermediate depths of over 800 feet on either side of the swell in the fiord bottom. The ridge near Miner's River rises 300 feet above the fiord bottom to the north, and 246 feet above that to the south; but its crest lies beneath 582 feet of water. As in the last case, it is difficult to interpret this shoal, which may be either an unremoved rock ridge due to basining of the fiord bottom or a glacial deposit. For the following reasons, it seems quite possible that it is the latter.

Miner's River, which is the outlet of Boot Lake and a smaller lake below it, is supplied by a stream from Pedro Glacier. At the time when Meares Glacier occupied the whole of Unakwik Inlet, Pedro Glacier was one of its tributaries and, as a much smaller glacier, should not have been able to erode as efficiently as the main ice tongue. In the case of the small glaciers of about the same size as Pedro Glacier, which occupied Jonah Bay and Siwash Bay, the inequality of glacial erosion resulted in these bays being left as submerged hanging valleys. The valley occupied by Miner's River, however, does not now have the hanging valley relationship to the main fiord, for the water is 384 feet deep at the head (Fig. 38), and 489 feet deep in the entrance of the cove, while opposite this cove, in the middle of the main fiord, the depth is only 582 feet. The hills near the outlet of Miner's River, however, separating Boot Lake from the main fiord seem to be morainic accumulations, and it is quite likely that Boot Lake is a former branch of Unakwik Inlet similar to Siwash Bay but separated from the main bay by morainic deposits. It is therefore considered possible that the Boot Lake valley is a submerged hanging valley of the usual kind, but that the discordance is not revealed because glacial accumulations were made directly opposite the mouth, shallowing the water from 850 feet, or thereabouts, to 582 feet.

Just inside the mouths of Jonah Bay and Siwash Bay are shoals, and Siwash Bay is deeper a mile or two back from the entrance (144 feet), than at its junction with the main fiord, where its depth is only 74 feet. In these cases also it has not been possible to determine whether the shoals are glacial accumulations or rock reefs.

The foregoing statement makes it clear that there is only one large accumulation in Unakwik Inlet which can be actually demonstrated to be, at least in part, a recessional moraine. This is the moraine bar rising nearly or quite to the surface in the main fiord opposite the mouth of Jonah Bay. The shoal to the north, near Miner's River, is possibly a recessional moraine while that to the south is more doubtful. If all three of these are morainic accumulations, however, they furnish interesting evidence of the intervals between halts in Meares Glacier as it retreated northward from Prince William Sound to its present position. It is five miles from the southernmost moraine to the one at Jonah Bay, and 4½ miles from there to the one opposite Miner's River which is 6½ miles from the present ice front, perhaps indicating that the glacier halted long enough to build these deposits at rather even intervals.

An alternate explanation of the irregularities in the bottom of Unakwik Inlet might be a rhythmic swing of erosion by a viscous ice body, with (1) a downward plunge at

the 882 foot basin opposite and below Brilliant Glacier, (2) another opposite and below Miner's River, (3) a third near Siwash Bay, (4) a fourth near Olsen Island (5) with unconsumed swells between where the ice rose after its plunge, perhaps because a cross current ran athwart its course where tributaries entered.

Above sea level, in Unakwik Inlet, deposits of glacial origin are inconspicuous, usually forming only a veneer of glacial till on some of the mountain slopes, and the coarse gravel fans, where the streams from Brilliant, Ranney and other glaciers enter the fiord.

Coast East and West of Unakwik Inlet

East of Unakwik Inlet the mouths of Wells and Long Bays (Pl. XCIII) are 1404 and 852 feet deep respectively. The fiords between, north of Fairmount and Glacier Islands, are 288 to 420 feet deep. Wells and Long Bays contain no tidal glaciers at present, but were undoubtedly sculptured profoundly by the former ice tongues in them, which were tributary to the ancient ice sheet of Prince William Sound.

Eaglek Bay, extending parallel to Unakwik Inlet just west of Pt. Pellew, is only 5½ miles in length. Eaglek Glacier, a small ice tongue, terminating half a mile from the head of the bay, was evidently once a tributary of the Prince William Sound ice sheet, for the sides of Eaglek Bay (396 feet deep) bear evidence of glacial erosion.

Esther Passage is a narrow channel between Esther Island and the mainland, connecting the east side of Port Wells with Prince William Sound. It has a length of about twelve miles, a width of from less than a quarter of a mile to a mile and a quarter, and is very sinuous in outline. The southward-trending eastern part of the passage has been notably modified by glaciation. Its depth, 142 feet, near where a glaciated valley comes in from the north, and 354 feet at the mouth, where it enters Prince William Sound, seems to be due to glacial erosion. Between these points is a place 552 feet deep, probably representing a glacial rock basin. The western part, which has a depth of 66 feet where it enters Port Wells, has not been much modified by glaciation.

College Fiord

General Description. West of Unakwik Inlet is Port Wells, which is the westernmost of the fiords entering Prince William Sound, and one of the largest fiords tributary to that body of water. Port Wells has two branches, College Fiord and Harriman Fiord, the latter coming in from the northwest, the former from the northeast. College Fiord, whose general trend is north northeast, has a total length of 24 miles, and varies in width from 2 to 3 miles, its walls being nearly everywhere precipitous. It has two indentations, both on the east side, the southernmost being a broad cove at the outlet of Coghill River 8 miles north of Port Wells, the second, Yale Arm, a tributary fiord 3 miles in length entering College Fiord 6 miles south of the head.

Harvard Glacier is at the head of College Fiord, Yale Glacier forms the head of Yale Arm. There are eight small cascading glaciers in the upper part of College Fiord, seven on the west side, and one on the east, and, of these, four descend the fiord wall to the water. In the lower part of College Fiord, there are five moderate-sized glaciers on the east side, including Dartmouth, Williams, Amherst, and Crescent Glaciers, all of which now end some distance back from the water. Beside these glaciers there are numerous

small ice masses on mountain slopes and in cirques, including Tommy and Cap Glaciers, south of Crescent Glacier, and scores of smaller, unnamed glaciers.

Previous Studies. The glaciers of College Fiord were seen by Whidbey[1] of Vancouver's party in 1794, Applegate[2] in 1887, Glenn,[3] Castner,[4] and Mendenhall[5] in 1898, the Harriman Expedition[6] in 1899, and Grant, Paige, and Higgins in 1905 and 1909.[7]

The most important of these visits were those by the Harriman Expedition and by Grant and Higgins. In 1899 Gannett of the Harriman Expedition made a general map of College and Harriman Fiords, showing the glaciers. The other members of the Harriman Expedition also took many photographs, and Gannett and Gilbert studied the glaciers as fully as their brief visit permitted. During Grant's several visits more photographs and descriptions were made, and a still more detailed map was made by Higgins.

In 1910 (July 15–25), the National Geographic Society's Expedition studied the glaciers and glaciation of upper College Fiord in some detail, making the contour map reproduced as Map 7 and the soundings shown on Figs. 43 and Plate CXXVI.[8]

Whidbey's description of conditions in College Fiord, quoted by Vancouver,[9] is of considerable interest. In June, 1794, he proceeded up College Fiord, from Point Pakenham at its southwest entrance, through "much floating ice." After going three miles "they met such innumerable huge bodies of ice, some afloat, others lying on the ground near the shore in ten or twelve fathoms water, as rendered their further progress up the branch rash, and highly dangerous. This was, however, very fortunately, an object of no moment, since before their return they had obtained a distinct view of its termination about two leagues further in the same direction, by a firm and compact body of ice reaching from side to side, and greatly above the level of the sea; behind which extended the continuation of the same range of lofty mountains, whose summits seemed to be higher than any that had yet been seen on the coast.

"Whilst at dinner in this situation they frequently heard a very loud rumbling noise, not unlike loud, but distant thunder; similar sounds had often been heard when the party was in the neighborhood of large bodies of ice, but they had not before been able to trace the cause. They now found the noise to originate from immense, ponderous fragments of ice, breaking off from the higher parts of the main body, and falling from a very considerable height, which in one instance produced so violent a shock, that it

[1] Vancouver, George, A Voyage of Discovery to the Pacific Ocean and Round the World, Vol. V, 1801, pp. 312–314.

[2] Applegate, S., Manuscript map reproduced by Davidson, George. The Glaciers of Alaska that are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc., Pacific, Vol. 3, 1904, p. 29 and map XI.

[3] Glenn, E. F., War Dept., Adj.-Gen. Office, No. XXV, 1899, pp. 19–21, and map (in pocket).

[4] Castner, J. C., *Ibid.*, pp. 190–191.

[5] Mendenhall, W. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 273–325, and Pl. XIX, A.

[6] Gannett, Henry, Nat. Geog. Mag., Vol. X, 1899, pp. 510–511 and map; Bull. Amer. Geog. Soc., Vol. XXXI, 1899, p. 354; Burroughs, John, Harriman Alaska Expedition, Vol. I, 1901, pp. 69–70; Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 81–89 and map.

[7] Grant, U. S. and Higgins, D. F., Glaciers of Port Wells, Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLIII, 1911, pp. 321–327; Tidewater Glaciers of Prince William Sound and Kenai Peninsula, Bull U. S. Geol. Survey (in preparation); Map in Bull. 443, U. S. Geol. Survey, 1910, Pl. II.

[8] Martin, Lawrence, The National Geographic Society Researches in Alaska, Nat. Geog. Mag., Vol. XXII, 1911, pp. 551–554, 556–560. Some Features of Glaciers and Glaciation in the College Fiord, Prince William Sound, Alaska, Zeitschrift für Gletscherkunde, Band VII, Heft 5, 1913, pp. 289–333.

[9] Op. cit., pp. 312–314.

was sensibly felt by the whole party, although the ground on which they were was at least two leagues from the spot where the fall of ice had taken place."

Harvard Glacier. Harvard Glacier, at the head of College Fiord, is an ice tongue certainly over 8 miles in length and probably at least twice as long. It heads on a divide at the crest of the Chugach Mountains and from the same pass Matanuska Glacier flows northward.[1] Harvard Glacier has a width of a mile to a mile and a quarter in the portion near the terminus and is fed by six good-sized tributary glaciers in the lower 6 miles, while many other tributaries enter above from cirques east of Mt. Grant.[2] The lower tributaries are all cascading glaciers, the only ones named being Radcliffe, Eliot,[3] and Lowell [3] glaciers. These cascading tributaries have an average slope of 165 to 2400 feet to the mile in their lower course, while the average slope of the lower 4½ miles of the main ice tongue is only 700 feet per mile. Harvard Glacier ends in a tidal ice cliff 350 feet high in 1899, the eastern part being lowest in 1910, as in 1899. That the glacier end is not afloat is proved by the fact that a half mile from the terminus the fiord is only 636 feet in depth.

The surface of the main glacier (Pl. CXIV, A) is severely crevassed throughout its whole visible extent, and upon it are eight prominent medial moraines, most of which can be traced backward to a point where they enter as lateral moraines from the margins of tributary glaciers. There are also several subordinate medials near the east side of the glacier. The position of the medial moraine formed by the north lateral moraine of the Radcliffe Glacier, indicates that in 1910 the Radcliffe was almost as strong as the main Harvard Glacier above the junction; for this medial moraine reaches the terminal cliff of Harvard Glacier nearly in the middle, showing that at least in the upper layers the ice from Radcliffe Glacier compresses the Harvard stream to half of its normal width. This is also shown in a unique way by the cross-section of all of the medial moraines revealed on the precipitous front of Harvard Glacier. These moraines, whose rock fragments darken the otherwise-white face of the cliff from top to bottom, are not vertical in cross-section (Pl. CXIV, B), but are inclined at angles of from 70° to 22°. The fact that the top of each inclined moraine section is farther east than the bottom, indicates that the push of the Radcliffe Glacier which comes in from the west is responsible for this inclined position.

It is possible that the reason why the Radcliffe Glacier is able to dominate the larger Harvard ice tongue is because the Radcliffe has so much steeper slope than the Harvard. Its slope is not much steeper than the other tributaries like the Eliot Glacier, but the descent in a short distance is greater. It may be that this is only a superficial influence, that is, that Radcliffe Glacier flows out and rides upon the surface of Harvard Glacier because of some rather recent advance.

There is a broad lateral moraine on each side of Harvard Glacier, the two lateral moraines at the terminus being respectively the south laterals of the Radcliffe and Lowell Glaciers, while the lateral moraines farther north on Harvard Glacier are diverted successively and become medial moraines below the entrance of each of the tributaries.

[1] See Pl. I, Bull. 327, U. S. Geol. Survey, 1907.

[2] Named in 1910 for Prof. U. S. Grant of Northwestern University.

[3] Named in 1910 for Charles W. Eliot and A. Lawrence Lowell, the former president and the present president of Harvard University.

It is impossible to say how much Harvard Glacier may have retreated from 1794 to 1887 and 1898, for the maps are not detailed enough. Applegate's map, made in 1887, seems to indicate that the terminus of Harvard Glacier had about the same relationship to the cascading glaciers on the western side of College Fiord as when Gannett made his more accurate map in 1899. The glacier seems to have changed very little between 1899 and 1905. An advance took place, however, between Grant's first visit in 1905 and his second visit four years later,[1] estimated to have been half a mile on the western side of the glacier and quarter of a mile on the eastern side. Gannett's map, made in 1899, when compared with the Higgins map ten years later, shows clearly that the advance could not have been so great as this without great retreat from 1899 to 1905. In 1910 the junior author compared photographs taken in 1899, 1905, and 1909 with conditions in the field and is inclined to think there was little if any retreat from 1899 to 1905, and to place the advance between 1905 and 1909 at a much smaller figure than Grant does, probably not more than two hundred yards. Between 1909 and 1910 the continuation of this advance amounted to only 100 or 150 feet.

This recent advance is shown clearly in photographs, which bring out the forward movement of the east side, and the overriding of a narrow barren zone on the western edge. The inclined moraines of the terminus were also pushed farther over by an advance of Radcliffe Glacier at the same time. The advance in 1910 resulted in increased crevassing in the lateral moraines on either margin of Harvard Glacier near the terminus, and in the overriding and destruction of forest on each side.

This advance was actually in progress when we visited Harvard Glacier on July 17 to 22, 1910. One of the pronounced effects of the advance was the increase in wave work as the result of more active discharge of icebergs. By these icebergs waves a delta, built by the stream from Downer Glacier, a quarter of a mile from the glacier front was being cut away at a very rapid rate. A precipitous wave cut cliff in the gravels of this delta had a height of 40 feet in 1910, most of the cutting having taken place during the previous year.

At the western edge of Harvard Glacier (Pl. CXV), where the advance in 1910 seemed to be due to activity of Radcliffe tributary, a push moraine was being formed on the beach. Along the glacier margin, a short distance to the northward, the moraine was made up largely of an inextricable tangle of macerated fragments of trees and roots, mixed with soil, moss, peat, gravel, and till. There were well-developed peat rolls, and in places the push moraine was 15 feet high. One of the trees which has just been overturned was a spruce 12 inches in diameter and probably over 100 years old, indicating that the glacier had not previously advanced as far as the present stage for at least a century. On this side of the fiord as on the eastern side there has been an increase in activity of iceberg waves which has resulted in the cutting of little cliffs, and the uprooting of shrubs.

Cascading Glaciers of College Fiord—General Description. On the western side of College Fiord there are seven cascading glaciers, as well as a number of much smaller ice masses. These Mendenhall[2] described in 1898 as follows: "Numerous small glaciers, easily distinguished from the unconsolidated snow by the blue color of their fronts and crevasses, hung from the summits, or extended down the slopes."

[1] Grant, U. S. and Higgins, D. F., Bull. Amer. Geog. Soc., Vol. XLIII, 1911, p. 325.
[2] Mendenhall, W. C., Twentieth Annual Report, U. S. Geol. Survey, Part VII, 1900, p. 272.

All of these cascading ice streams were formerly tributaries to Harvard Glacier, and they have been detached from the main glacier as a result of its retreat northward past each successive cascade. In the same way a further retreat of Harvard Glacier for a mile and a half would transform Radcliffe Glacier into a separate cascading tongue.

Named in order from Harvard Glacier southward, these cascading ice falls are Baltimore, Smith, Bryn Mawr, Vassar, Wellesley, Barnard, and Holyoke Glaciers. On the eastern side of the fiord, opposite Baltimore Glacier, is Downer Glacier, which has rather recently ceased to be a tributary of Harvard. Smith, Bryn Mawr, Vassar, and Wellesley Glaciers have sufficient snow supply so that they still extend down the fiord wall and end in the water (Pl. C, B). The other cascading glaciers, with smaller snowfields, end some distance above the fiord,—1004 feet in the case of Baltimore Glacier and 950 feet in the case of Downer.

The slopes of these cascading glaciers average about 2500 feet to the mile, being very much greater in some parts than in others because the lower slope of the fiord has been much oversteepened by glacial erosion and because the upper parts of some of the glaciers are in hanging valleys. In each case the lower, or cascading portion of these glaciers is in a shallow valley, interrupting the slope of the fiord wall.

Several of the cascading glaciers have built terminal moraines and outwash fans in the fiord, giving rise to a narrow piedmont flat at the base of the western fiord wall. Vassar and Bryn Mawr Glaciers spread out upon this flat, so that below the ice cascade there is a gently-sloping terminus extending from the base of the mountain to the sea. Wellesley Glacier formerly had such an expanded foot, but has retreated so that a good-sized cove interrupts the narrow coastal plain, the glacier ending three quarters of a mile back from the main fiord. Within the cove the water is 48 feet deep.

From Downer, Baltimore, Barnard, and Holyoke Glaciers, which end high above the fiord, large streams descend in foaming waterfalls, building deltas where they enter the fiord. A similar stream descends the fiord wall from a minor lobe of Vassar Glacier, which ends 2000 feet above the fiord.

Downer and Baltimore Glaciers. This pair of ice tongues which would be the first ones to reunite with Harvard Glacier if there should be general advance, have not changed greatly since 1899, the period of photographic observations. Downer Glacier [1] is a clean, white ice tongue with no moraine. Beyond its end there is an extensive barren zone 950 feet above sea level, which shows that there has been retreat in recent years. Indeed, fully a third of this barren zone has developed since 1899, and part of it may be glacial ice, mantled with ablation moraine.

The stream from Downer Glacier descends through a steep-sided rock gorge, below which is the torrential delta, already referred to. Upon its surface were shrubs, none over 6 or 7 years old, indicating that there had been no great activity of delta building since 1902 or thereabouts.

Baltimore Glacier, which is fed from snowfields that coalesce with those supplying Smith Glacier, is also a clean white ice mass, with no moraine except small medials near the terminus. Half a mile from the end it bifurcates, the north lobe being larger and and ending lower (1004 feet) than the south lobe. A barren zone about its terminus indicates that the glacier is less extensive than it was a few years ago; but while Radcliffe and other cascading glaciers to the south advanced between 1909 and 1910, there

[1] Named in 1910 for the Milwaukee-Downer College for Women, in Wisconsin.

is no evidence of advance in the Baltimore Glacier, that could be detected by comparison of photographs. There were shrubs at the very edge of the ice, but from our viewpoint at sea level it was impossible to tell whether any had been overridden. Along portions of the margin ice blocks were sliding down the steep slope and across the barren zone. We were inclined to suspect advance in progress in July, 1910, but could not be certain.

Smith Glacier. Smith Glacier, which has a length of about 3½ miles and a width of from 1800 to 3200 feet, is fed by two large and two small tributaries which head in cirques 4000 to 5000 feet above the fiord. The glacier descends at the rate of 2200 feet to the mile and is severely crevassed. The lip of its hanging valley is at a level of 1395 feet, and there are subordinate cascades higher up. There are broad dirt-stained areas on either margin, but no definite lateral moraines. There is a pronounced medial moraine, formed by two lateral moraines uniting at the spur 2387 feet above the fiord, and another from the 3000 foot spur on the east wall. The terminus is a precipitous ice cliff rising about 100 feet above the sea. At sea level just north of the middle of this ice cliff, rock ledges are exposed beneath the glacier.

Gilbert states that in 1899 Smith Glacier reached "the fiord three or four miles from the Radcliffe, and is of the same order of magnitude. Fed by several tributaries among the crests of the range, it gathers in a high mountain valley, and then descends in magnificent cascades down the mountain front to the sea. In the last part of its course it has scarcely any valley, the outer surface of the ice being practically flush with the face of the mountain; and there is no flattening of its profile, as it reaches the water. Though its lower slope is so seamed by crevasses as to exhibit a mere congeries of pinnacles, two lines of medial moraine are distinctly traceable, each partitioning off a fourth part of the ice stream at the side."

This description and the photographs made by the Harriman Expedition, show that Smith Glacier changed very little between 1899 and 1909. One of these photographs (E. H. H., 81) shows a narrow barren zone on each margin of the glacier extending up the mountain slopes to about 1400 feet. In a photograph taken by Grant on July 1, 1909, the same barren zone apparently shows on the north side of the glacier. In July, 1910, Smith Glacier was actively advancing, apparently having commenced since Grant's visit the year before; and no barren zone was left around its borders. On the south side at sea level, the glacier had spread to the edge of the barren zone, where, 15 or 20 feet from the glacier, an earlier lateral push moraine crossed a marginal stream, down which small ice fragments were floating. Higher on the mountain slopes the ice was advancing into the forest and destroying shrubs.

The north side of Smith Glacier had spread farther than the south side, and at the time of our visit was just beginning to destroy the alder along the margin, except at sea level where it was advancing across a narrow zone of bare rock. Along the advancing margin the alders were being destroyed in three ways,— by actual overriding of the spreading glacier, by stream encroachment, and by ice-block avalanches which rolled some distance out into the forest (Pl. CXVI), knocking down and breaking off shrubs and removing their bark 6 or 8 feet above the ground. By the advance a marginal stream had been diverted some distance northward into the forest.

It was impossible to tell exactly how much the tidal terminus of the glacier had moved forward since 1909 but there was undoubtedly several hundred feet of advance, accompanying the spreading on the north and south margins. Plate CXVII, A, from the north

side shows a flat tidal terminus extending a short distance out into the fiord, where Gilbert says there was none in 1899. The extreme southern edge of the ice cliff was a black, crevassed precipice, the lateral moraine of 1899 having been pushed forward into the sea.

Bryn Mawr Glacier. Bryn Mawr Glacier (Pl. CXVIII), the largest independent cascading glacier in College Fiord, is over 3 miles long and varies in width from $\frac{3}{4}$ of a mile to a mile. It is fed by two large tributaries which unite 2040 feet above sea level, coming from cirques at the level of about 4500 feet. Below the lip of the main hanging valley, 1341 feet above sea level, the slope of the glacier averages 3700 feet to the mile. Above this level the slope flattens to the junction of the tributaries, each of which plunges down from a secondary hanging valley at a level of about 2500 feet. At the base of the steepened valley slope the glacier forms a slightly expanded piedmont bulb, and the lower $\frac{3}{8}$ of a mile has a very low, flat surface grade. As in the Smith Glacier, there are no well defined lateral moraines, but a strong medial moraine extends from the junction of the tributaries to the sea. The glacier discharges more icebergs than any of the other cascading tongues and terminates with a precipitous cliff over 100 feet high.

Gilbert states that the Bryn Mawr Glacier "next south of the Smith, is somewhat larger. Its two main branches, gathering in mountain valleys not well seen from the sea, became visible in twin cascades, and then, uniting their streams, make a second leap to the sea. As tide is reached, there is a tendency to flatten the profile, and the central portion of the stream becomes nearly or quite horizontal for a few hundred feet before breaking off in the terminal cliff."

This shows clearly that the expanded lower portion with a flat foot, was already developed in 1899 and photographs by Grant and Paige prove this condition to have continued in 1905 and 1909. There was a narrow barren zone on either margin in 1899 and probably in 1905. Grant has stated that "a comparison of the photographs taken in 1899 with those taken in 1909 indicates that the glacier was farther advanced at the latter date and that its front (especially the southern half of the front) deployed more widely on the shallow bottom of College Fiord. A photograph taken in 1905 and an impression four years later indicates that the glacier was less advanced at the earlier date, and that it was then (1905) at approximately the same position as in 1899. Any close estimate of the actual amount of this advance (as recorded in the photographs taken in 1909) is impracticable from the data at hand, but it is probably as much as 500 feet." His photographs show an extensive gravel beach still bordering the southern half of the ice front in July, 1909.

There was very considerable advance between July, 1909, and July, 1910, for a photograph from Grant's site on College Point shows a lateral spreading of several scores of feet on both north and south margins of the glacier. That there has also been pronounced forward movement is evident by comparing the conditions in July, 1910, with a photograph taken by the Harriman Expedition in 1899. In this interval, the Bryn Mawr Glacier had advanced several hundred feet, most of the advance apparently taking place during 1910.

On each side of the glacier a small stream emerges from the ice, and at the time of our visit the borders of the glacier were encroaching on these stream courses. All along its northern margin the Bryn Mawr Glacier was advancing into the forest, where it was killing spruces up to 5 inches in diameter, suggesting that the glacier had not been so large for

a half century or thereabouts. Sometimes as much as 25 feet outside the area where trees were being overwhelmed, shrubs were crushed and trees barked and broken by blocks of ice that had tumbled from the glacier margin.

The south side of Bryn Mawr Glacier showed active advance at the time of our visit, the margin of the glacier having a push moraine made up of gravel on the beach and alluvial fan, and of turf and soil farther back from the shore. The site of a push moraine of 1899 was completely overridden and the new push moraine, 6 to 10 feet high, was filled with trunks of overwhelmed shrubs, many of them still retaining the leaves of 1910 growth. In some places the turf outside of the push moraine had been shoved up into low arches by the advancing glacier. Willows and alders that were being overwhelmed were from 10 to 15 years old, and in one case possibly as much as 25 years old. There was no barren zone left on the south side of the glacier, the advance having entirely covered it.

This advance had displaced the south marginal stream so that it was flowing through the bushes south of its previous course and carrying them down-stream. At one point a lobe of the ice which was swept by the stream had been undercut until a small avalanche of ice blocks was formed and these blocks were floating down the stream toward the sea, coming from a point on the glacier margin nearly a quarter of a mile from the fiord. Near the beach the drainage was obstructed by the ice advance and several small pools were formed between the advancing ice front and an older terminal moraine. At several points along this southern margin of Bryn Mawr Glacier the upper layers of the ice had been thrust-faulted forward over the lower layers so that splinters of ice projected into the forest, as was the case on a larger scale at Columbia Glacier.

The area on the southern border of Bryn Mawr Glacier which is being overridden, is part of a crescentic terminal moraine, with knobs and basins, built by the glacier at a time of much greater expansion, and now grassed over and partly covered by alders and spruces, some 65 years old. It is, therefore, at least that long since this older terminal moraine was built.

Vassar Glacier. Vassar Glacier differs from Smith and Bryn Mawr in having a well-defined bulb at the lower end. The glacier heads in cirques 4000 to 5000 feet above sea level, and is about 2½ miles long, a mile wide at its snowfield, and three-eighths of a mile wide where it plunges over the lip of its hanging valley at a level of 1598 feet, above which is a second cascade at a level of a little more than 2200 feet. After cascading down the hanging valley lip to the fiord level, with a slope of 2500 feet to the mile, the glacier extends eastward a half mile further with a much flatter grade and with a lobate lateral expansion. This glacier with its three portions, a moderately-sloping upper part, a cascading middle section, and a flattened terminus, reminds one of the Rhone Glacier in the Alps, except that it terminates in the sea, while the Rhone Glacier ends in a valley on the land. As already stated, a minor lobe of the Vassar Glacier terminates 2000 feet above the fiord and sends a stream down the fiord wall.

Vassar Glacier has no lateral moraine on the south side, but there is a weak one on the north, and one medial moraine, starting below the lip of the hanging valley, but not from a lateral moraine, as there are no noteworthy tributaries. The whole of the lobate lower portion, however, is mantled with thick ablation moraine, which extends up the southern margin of the cascading portion to a height of 1100 feet, but on the northern margin ends at an elevation of about 300 feet. In contrast with the other

cascading glaciers, therefore, Vassar Glacier is the least attractive, because the whole of its lower surface is dark and moraine-covered. On both the north and south sides, streams emerge from the ice about a half mile back from the terminus and flow near the ice edge, building an alluvial fan on each side.

The tidal terminus of Vassar Glacier is not a perpendicular white cliff of clean ice, as in the other tidal cascading glaciers, but a low sloping margin, mantled with rock débris, and similar to the terminus of the Malaspina Glacier at Sitkagi Bluffs. As a result of a slight advance which was in progress in 1910, a part of the northern portion of the glacier was acquiring a more precipitous cliff and was shedding the débris mantle, so that the ice was revealed in places. Even here, however, in contrast with the glistening white cliffs of the adjacent glaciers, the ice in sight was dirty and marked with wavy horizontal lines of englacial material. Few, if any, icebergs are discharged from this glacier.

Gilbert describes Vassar Glacier in 1899 as a cascading glacier "parallel to the Smith and Bryn Mawr and exhibiting a similar series of cascades, but of smaller size and less direct in its course. It is cumbered, especially in its lower part, by rock débris, and close inspection was necessary to determine the fact that it is actually tidal."

It was, therefore, not essentially different in 1899 and 1910, and photographs by Grant show this to have been the case also in 1905 and 1909. It is not smaller than the Smith Glacier, though it is smaller than Bryn Mawr. The establishment of the facts of the débris-covered lower end and the barely tidal condition in 1899 are of importance, for these remained the same until 1910 when a change was commencing. At the time of our visit the glacier touched tidewater along the whole portion of the front between the flanking alluvial fans, but with a low, sloping moraine-veneered margin along the southern half, and with a low, dirty, nearly vertical cliff in the northern half (Pl. CXIX).

Photographs show that there were barren zones on each side of the glacier in the cascading portion in 1899. There was sufficient advance of Vassar Glacier before 1910 so that portions of the barren zones were covered. That on the north side near sea level was not completely overridden by July 21, 1910, but higher on the fiord wall it was almost covered. A great deal of the barren zone on the south side was still visible except in the bulb portion near sea level.

On the northern margin the glacier was advancing, but there was no thrusting forward of the ice itself, the advance being manifested chiefly by a slight folding, ridging, and crumpling in the moraine-covered hill along the northern margin, whose barren condition suggests that it contains a buried ice block.

The southern edge of the glacier at sea level was also obviously advancing, but no ice was seen beneath the ablation moraine cover and the advance was manifesting itself principally by the commencement of crevassing in the moraine-covered bulb and by the sliding down of this morainic material along the glacier margin, where willows and alders were being buried. Near sea level the glacier extended right up to the forest which included mature spruces.

High above sea level, the ice edge had advanced over part of the zone of bare rock that had existed there since 1899, but it was not all covered. Ice blocks were sliding down and accumulating in a talus at one point, suggesting that the advance was still in progress.

Vassar Glacier had been stagnant and motionless long enough, not only for thinning

until the débris carried by the ice had accumulated on the surface, forming the ablation moraine, but also long enough for the growth of a few shrubs and of moss and flowers upon the moraine-covered surface. This vegetation upon the bulb was very sparse, however, indicating that melting had gone on fast enough to interfere with plant growth. If the advance of Vassar Glacier which began in 1910 continues long enough, crevasses will doubtless break up the lower bulb, so that the ablation moraine will disappear and the lower end of the glacier be transformed to a white mass similar to Bryn Mawr Glacier, instead of the present dirt-covered terminus.

Wellesley Glacier. Wellesley Glacier is a very symmetrical single, cascading ice tongue about ½ mile wide (Pl. CXX). It descends the fiord wall from small tributaries in cirques at elevations of over 4000 feet, reaching the water's edge, but not expanding in a piedmont bulb. It descends from a fine hanging valley (Pl. CXXV) at an elevation of between 1700 and 1800 feet, the slope below the lip exceeding 28°. There are inconspicuous lateral moraines and the suggestion of a weak medial moraine, close to the northern margin.

It is evident that not long ago Wellesley Glacier extended at least ¾ of a mile further than now, having then a form very much like that of Vassar Glacier. In retreating from this more advanced position the glacier has left a narrow barren zone along the north and south margins. The site of the central part of this former bulb is occupied by a moraine-bordered cove with water 21 to 129 feet in depth, into which a spit extends from the south side, marking the site of the former terminal moraine. At the head of this cove Wellesley Glacier ends with a vertical ice cliff, beneath the northern and southern margins of which rock ledges show.

Gilbert speaks of it in 1899 as a cascading glacier which "flows with gentle grade through a mountain trough joining the fiord at right angles and then cascades into the sea, into which it plunges without notable modification of profile." This description and the 1899 photograph of the Wellesley Glacier show that it was then essentially as in 1910, and that there was then as now, a broad terminal and lateral barren zone. Although at the time of our visit there was a very much larger barren zone around the glacier terminus than around any other ice tongue in College Fiord, it was then actively advancing, and the northern and southern margins had partly covered the lateral barren zone previously exposed.

On the north side, part of the glacier margin was bordered by a push moraine from 5 to 8 feet high. The advancing edge was lobate, and parts of the barren zone on the mountain slopes were completely covered. Both the ice and the push moraine were overwhelming alders 5 to 6 years old, and broken ice blocks were rolling down the edge of the glacier.

The southern edge of Wellesley Glacier was also bordered by a new push moraine in which were torn willow and alder bushes, some of them 15 years of age. The edge of the ice did not extend across the barren zone, which contained shrubs 8 years old, so that no trees were being overwhelmed by the advancing glacier, although annual plants of 1910 growth were being overturned and buried. The south barren zone has a morainic topography with linear crevasse deposits, knobs, and basins, the latter occupied by pools.

The barren zone in front of the glacier terminus, including the land portion of the former bulb, has evidently been covered by glacial ice within ten or a dozen years, for

the oldest shrubs growing in this barren zone were from 8 to 10 years old. Outside this morainic area are mature alder thickets and some spruces.

Barnard, Holyoke, and Smaller Glaciers. Barnard and Holyoke Glaciers, smaller cascading glaciers than most of those toward the north, end a thousand feet or more above the fiord. Gilbert alludes to them as "small glaciers occupying alcoves on the mountain front but ending far above the water." South of them are three small cliff glaciers, resting on ledges at the base of Mt. Emerson. There are also small ice masses between several of the cascading glaciers to the north.

Barnard Glacier is a clean, crevassed ice tongue, with no medial or lateral moraines. It heads in a large cirque and connects with an irregular ice mass on a shelf to the north. There is a bare rock slope at the end of the glacier which has small terminal lobes and terminates on the lip of its hanging valley. Two streams descend from the terminus. Mature spruce forest extending from the fiord well up toward the glacier terminus indicates that Barnard Glacier has not descended much farther toward the fiord for a century or more; but a barren zone between the ice and forest, present in 1899 as well as in 1910, proves that it has been retreating in recent years. Between 1899 and 1910 there was an advance of the south lobe down the lip of the hanging valley, and a slight advance of the north lobe, while between the two lobes an ice block talus was formed. We are inclined to believe that the advance was still going on in 1910.

Holyoke Glacier, which heads in a large cirque and is fed by two glaciers from a small cirque on the south, has two weak medial moraines extending from spurs between its south tributaries. There are no lateral moraines, but there are a few morainic patches along the terminus. The glacier is longer than Barnard, and extends out over the lip of its hanging valley. In 1910 the glacier nowhere extended to the borders of its barren zone, and the mature spruce forest between the small terminal barren zone and the fiord demonstrates that it has not extended beyond this barren zone for a century or more. No distinct signs of recent advance were seen.

The cliff glaciers on the ledges south of Holyoke Glacier present no unusual features. There is every reason to believe that they are self-sustaining, and in one case a projecting lobe extended down from the shelf.

Yale Glacier. Yale Glacier has a known length of seven miles, and is probably over twice that length. Its width varies from a mile and a quarter to two miles. In its lower portion the glacier slopes at the rate of 600 to 700 feet per mile, attaining an elevation of 2500 feet three miles and a half from the front, and ascending gradually to 6000 or 7000 foot cirques east of Mt. Glenn.[1]

The Yale Glacier, though wider at the terminus, is probably not as long as the Harvard. It terminates in the Yale Arm of College Fiord, with an unusually irregular front, the south side of the glacier extending $1\frac{1}{4}$ miles further down the fiord than the north side. This terminal cliff is between 200 and 300 feet high. There are well-marked lateral moraines, but no medial moraines (Pl. CXXIV, A). In the absence of a number of large tributaries which supply quantities of ice, and lateral moraines that become medials, Yale Glacier differs very decidedly from Harvard Glacier. The distant tributaries are rather small glaciers cascading from extensive névé fields on the mountain slopes. In the lower part of the glacier there are not many tributaries on the northwestern side, contrasting with a considerable number which descend on the southeastern

[1] Named in 1910 for Lieut. E. F. Glenn of the U. S. Army.

side from the snowfields and cirques about Mt. Castner [1] and other mountains between College Fiord and Unakwik Inlet.

Applegate's map of Yale Glacier in 1887 shows it terminating somewhere between College Point and the present stand; but since he did not approach the glacier nearer than 12 miles, about the same as Vancouver did in 1794, it is impossible to draw any conclusions from these early maps concerning the behavior of Yale Glacier prior to 1898.

From the photograph (Pl. CXXIII) taken by Mendenhall in the latter part of April, 1898, it is evident that Yale Glacier was in almost exactly the same position as at present. The rock ledges were exposed beneath the middle of the ice front and the eastern half exhibited "the rough pinnacled front of a still-advancing stream. Its western front is of dead-white ice."

On this same occasion Castner went on snowshoes some distance up the margin of the Yale Glacier, which may perhaps show that it was then less severely crevassed than in 1910 when traveling upon its surface was impossible. But our visit was at a later season when the snow had disappeared.

Glenn gives a vivid description of Yale and Harvard glaciers as he saw them in 1898 from College Fiord. "The day was dry and clear. Directly in our front was the most imposing sight we had yet seen—I might add more imposing than any we saw during the season. Glistening in the sun were two large glaciers, which we named the 'Twin Glaciers,' [2] the pair being separated by a short ridge or hogback that runs down to salt water. In front of the one on our right the sea ice extended for over 3 miles, while in front of the other this sea ice extended at least twice that distance. This ice was covered with snow several feet in depth. We soon discovered that it would bear up the weight of a man and that we could make no headway against it with the boat. Each of these glaciers is what is termed 'live' or 'working' glaciers. The front of each was an almost perpendicular mass of ice, from which immense pieces were constantly breaking off and falling into the sea with a great roaring noise, due principally to the action of the tides."

The map of Yale Glacier by Gannett, the description by Gilbert, and the photographs by other members of the Harriman Expedition in 1899 show clearly that in all major features of position Yale Glacier had at that time assumed the general conditions which prevailed in 1910. The rock ledges beneath the center of the glacier are seen in the photographs, and Gilbert indicates that there were barren zones at the margins. He explains the dirty ice near the northwestern side of the glacier as follows: "A blackening, west of the middle, by glacial drift suggests that a rock knob may lie near the surface, ready to develop into a nunatak or island if the glacier shall diminish."

The photographs and descriptions by Grant in 1905 and 1909, and the map by Higgins in the latter year show that Yale Glacier maintained essentially the same conditions during the decade following 1899. The narrowness of the barren zones in 1909 leads Grant to conclude that the eastern margin was slightly farther advanced in that year than in 1898 and 1899.

In 1910 Yale Glacier was advancing strongly. By July 15, the southeastern edge had advanced 750 feet beyond its position in 1899, as was shown by comparing conditions

[1] This peak and the adjacent glacier were named in 1910 for Lieut. J. C. Castner of the U. S. Army.

[2] Now called Yale and Harvard Glaciers.

seen from Photo Sta. P., with a photograph (Pl. CXXI) by Curtis.[1] The amount of advance was accurately determined by Mr. Lewis, our topographer. That the greater part, if not all, of this advance took place after July 1, 1909, is demonstrated by photographs by Grant from Station I (Pl. CXXII). The central portions of the glacier also advanced somewhat, covering part of the ledges. The southeastern ice cliff was highest in 1910, the fresh-looking cliffs west of the rock ledges next highest, and the northwestern margin lowest of all.

The advance had covered most of the barren zone along the southeastern margin, though some still remained in places. Near the terminus the entire marginal barren zone was covered by ice, and bushes were being overridden, but in front of the glacier there were rock slopes bare of vegetation. Bordering part of the glacier margin was a push moraine of till, bowlders and wood, and there were also some peat rolls. The presence of recently overturned bushes lying on the winter's snow, proves that the advance was in progress during the spring of 1910.

The northwestern margin of Yale Glacier was also advancing rapidly. At the water's edge the glacier terminated on beach and alluvial fan gravels in part undisturbed; but above tide level there was a gravel push moraine including some tree trunks. Farther back along this margin practically all the former barren zone had been overridden, and the glacier was destroying thickets of alder and willow bushes from 10 to 33 years old. The relationship to snow banks proves that part of this advance had taken place during the preceding autumn or early winter; but on July 16, 1910, the advance was still in progress, overwhelming a dense growth of alders and willows, and at a few points rolling up small arches of peat.

Castner Glacier, the southernmost tributary of Yale Glacier, with its pronounced medial moraine and broad lateral moraine, showed no particular change from 1899 to 1910. Higgins' map has this glacier disconnected from the Yale in 1909, but Grant's photographs of that year show this to be an error. This lack of change indicates that the advance of Yale Glacier was not caused by this tributary. The feeders farther back on the east side may have participated in it, for they rise in snowfields which also feed the active northern arm of Meares Glacier in Unakwik Inlet.

Amherst and Adjacent Glaciers. This group of glaciers, on the southeast side of College Fiord has not been studied in detail by any of the expeditions that have visited the region, nor seen from nearer than the eastern edge of the fiord. Amherst Glacier is the largest of the group and both it and Crescent Glacier were shown on a map by Applegate in 1887 and by Gannett in 1899. Williams, Dartmouth and a small glacier to the north were indicated on the Grant and Higgins map of 1909, as were the Tommy and Cap Glaciers south of Crescent Glacier.

Amherst Glacier is about 4½ miles long and a mile wide and descends from large cirques near Unakwik Peak. Crescent Glacier is of about the same length but is only about a quarter of a mile wide. Dartmouth and Williams Glaciers are also small. According to Applegate's map, which was made from so near that it is probably correct, Amherst and Crescent Glaciers ended a little nearer the fiord in 1887 than at present and were united.

Gilbert states that in 1899 "Amherst Glacier was passed by the ship at some distance, and its features are known chiefly through the photographs secured by Merriam. It is fed by névés in full view from the fiord, and approaches the sea in a short, broad stream

[1] Harriman Alaska Expedition, Vol. III, 1904, Fig. 43, p. 83.

which at first descends steeply and afterwards more gently. The habit of the lowland lying between the glacier and the ocean indicates that it is built of morainic material. Near the sea is a belt of timber, but this is separated from the ice by a barren tract similar to that about Davidson Glacier. A barren zone several hundred yards broad is seen to flank the glacier on the southwest, and a similar zone borders its companion Crescent Glacier. These features doubtless indicate shrinking in modern times, the change having been of moderate amount, although greater than in the case of LaPerouse and Columbia Glaciers. The Crescent is comparatively narrow, and approaches the sea with a higher grade. A curve in its trough conceals its upper course."

When we saw it in 1910 (Pl. CXXIV, B) Amherst Glacier was little crevassed and had a broad belt of ablation moraine upon the terminus, no changes being detected by comparison with Merriam's 1899 photograph. The Crescent Glacier also seemed unchanged, and from a distance no signs of unusual activity were seen in Williams, Dartmouth, and Tommy Glaciers. Cap Glacier is typical of many névé fields upon mountain slopes, being thin, severely crevassed, and intermediate in character between the snowfield and the valley glacier, though most resembling the former.

Glacial Erosion—Features above Sea Level. Glacial erosion has had a profound influence in producing the present straight, steep-walled depression of College Fiord. Evidence of glacial erosion in the form of striae on rock ledges, in roches moutonnées forms, in the absence of spurs on the valley walls, and in the oversteepened lower portion of the fiord walls is everywhere apparent. There are also cirques of various sizes, a large number of them containing small glaciers, as between Vassar and Bryn Mawr Glaciers, between Baltimore and Radcliffe Glaciers, etc.

There are many differences in the degree of glacial erosion, judging from the valley-wall features. For example, the south side of Cresent Glacier valley shows an undercut cliff, evidently due to long-continued glacial erosion on the outer side of a curve, as in a river. The eastern wall of College Fiord, especially between College Point and Downer Glacier, and just north of Coghill River, is remarkable for irregularities above sea level in contrast with the smoother western side. The eastern side has many uneroded spurs, irregular hummocks and reefs; the western is remarkable for its smoothness and simplicity of contour, for its many roches moutonnées with striae and glacial grooves parallel to the trend of the fiord, and for the bare, smoothed, nearly horizontal rock ledges high upon the fiord wall. North of College Point the expanded Harvard Glacier seems to have hugged the western shore and to have eroded here more efficiently, despite the presence of many tributaries from the west whose thrust would tend to push the glacier over toward the eastern side. But it is very probable that the expanded Yale Glacier swept over College Point and by its great volume overcame the influence of this thrust.

There are many hanging valleys above sea level (h, Pl. CXXVI) and as Gilbert has already noted, the cascading glaciers on the western side of the fiord descend out of hanging valleys. There are also hanging valleys out of which the glaciers do not now cascade. The ascending altitudes of these hanging valleys, whose levels as measured in 1910 agree except in minor particulars with Gilbert's computations in 1899, and the fact that there are two steps, as noted by Gannett, have been explained by Gilbert,[1] as follows, the relationships being shown in Fig. 41.

[1] Gilbert, G. K., Harriman Alaska Expedition, Vol. 3, 1904, pp. 175-176.

"My attention has been directed by Gannett to the fact that several of the cascading glaciers make two leaps, and that there is a certain amount of harmony in the spacing of the falls. When the region shall have been thoroughly studied it is possible that the interpretation of these correspondences may develop a special chapter in the history of the ice retreat.

"With the aid of a series of photographs made by Merriam, I have computed the approximate heights of the more important cascades, as follows: Wellesley, 1700 feet; Vassar, 2200; Bryn Mawr (trunk), 1300, (left branch) 2700, (right branch) 2500; Smith, 1250, 1700 and 2600; Radcliffe, 1800 and 3500. When these are platted to scale in their proper vertical and horizontal relations they fall into two series, descending

FIG. 41. HANGING VALLEYS ASCENDING NORTHWARD IN COLLEGE FIORD AFTER G. K. GILBERT.

southward from the head of the fiord. Making some allowance for the greater volume of the side glaciers when the trunk glacier filled the fiord, I have indicated the profile of the trunk glacier by a dotted line (AB). The inclination of this line from the horizontal is about 2°, or one in twenty-five. Its height above tide ranges from 2800 to 4800 feet, and it indicates a thickness of ice exceeding these figures by the depth of the fiord, whatever that may be. In the line of Gannett's suggestion, a second tentative profile (CD) is drawn in similar relation to the crests of the lower series of cascades.

"The depth of ice indicated by the hanging valleys is somewhat less than that which would be inferred from the rounding of projections, and it seems probable that the epoch during which the hanging valleys received their principal sculpture was not the epoch of maximum glaciation."

Submarine Topography. The soundings made in College Fiord in the summer of

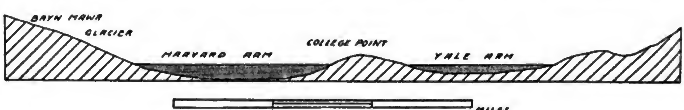


FIG. 42. NATURAL SCALE CROSS-SECTION OF COLLEGE FIORD.

1910 enable us to carry the discussion of glacial erosion farther than was previously possible (Pl. CXXVI). The depth of water near the fronts of Harvard and Yale Glaciers, 636 and 282 feet respectively, proves that neither of those glaciers is afloat, and that they must now be eroding on the fiord bottom. The line of rock ledges visible under the edge of Yale Glacier shows that in the past erosion has been less active in mid-fiord than on either side.

The water along the axis of College Fiord varies in depth from 174 to 804 feet, and the cross-sections of the fiord show that it has the typical U-shape (Fig. 42) below sea level. The longitudinal slope away from Harvard Glacier is interrupted near the terminus of the Smith Glacier where the depth decreases to 306 feet, by another swell

(174 feet) northwest of Coghill River, and by a gradual shallowing of the water from that point to Port Wells.

As has been stated in previous discussions of the 1910 soundings, we are in doubt where to infer submerged moraines and where uneroded reefs, developed in connection with the basining of the fiord bottom by glacial erosion. The basin whose deepest point (804 feet) is opposite Wellesley Glacier seems to us probably to be a rock basin of glacial erosion. Its bottom lies 630 feet lower than the swell at its lower end, and 462 feet lower than the swell at its upper end. An apparently adequate explanation of excessive erosion here would be the more rapid motion of the ice when the expanded Harvard Glacier received Yale and the cascading glaciers as tributaries. Moreover, in this upper part of the fiord there may have been repeated ice advances, with consequent periods of glacial erosion, while the lower part of the fiord was not scoured so frequently, if indeed it was occupied by an advancing glacier more than once.

An alternate explanation of the relative shallowness in the part of the fiord south of Coghill River (Fig. 43) would be that it was originally more deeply eroded, but has been filled by glacial deposits, including not only deposits from the retreating glaciers of upper College Fiord, but also by stream-borne deposits from the Amherst, Crescent, Williams

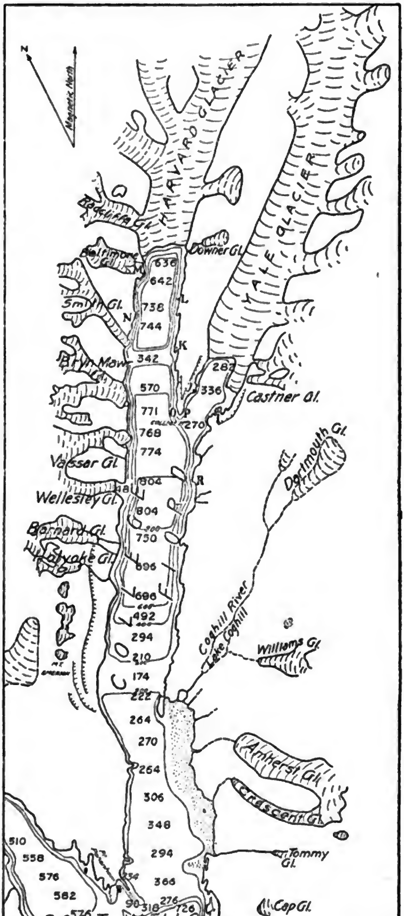


Fig. 43. Submarine Contours of College Fiord. Outline of coast and glaciers after Grant and Higgins.

and Dartmouth Glaciers, the streams from which are still carrying débris into the fiord. The eastern part of the fiord near these glaciers, however, is very shallow and the abundant reefs rising to the surface here are in some cases rock ledges, proving that there has not been such profound erosion as along the western side and also that there have not been extensive glacial deposits above the surface. The greater glacial scouring was apparently on the western side of the portion of College Fiord where the main current of the expanded trunk glacier was probably pushed by the eastern tributaries. Deposits might also have been made here by the Amherst and other tributaries before they retreated across to the shallow eastern portion of the fiord. A perfectly rational hypothesis to account for all the irregularities in the bottom of College Fiord would be to attribute them to glacial erosion without significant deposition. All of the existing basins and swells might be explained by a rhythmic alternation of plunging and rising related to the incoming of tributary ice streams. There is a basin below Harvard Glacier terminus and a swell not far above College Point near Smith Glacier. There is a basin south of College Point where the expanded Yale Glacier entered, below which is the swell west of Coghill River. Below this the water deepens nearly to Point Pakenham, perhaps under the influence of the incoming Dartmouth, Amherst, and adjacent glaciers.

The places of relatively shoal water at the southern end of College Fiord, discussed in the next chapter, seem to be clearly of the moraine bar type, though we have no clear proof that they are not built upon a broad rock swell, just above the entrance of the former incoming Barry Arm Glacier.

Besides the visible hanging valleys above tide water (Pl. CXXV) there are several submerged hanging valleys (Pl. CXXVI). The cove in which Wellesley Glacier now ends, and which has a maximum depth of about 48 feet, hangs 756 feet above the bottom of the main fiord. The arm of the fiord terminated by Yale Glacier hangs 498 feet above the main fiord (Fig. 42). This submerged hanging valley shows clearly that Harvard Glacier has always been the chief tributary of the College Fiord ice tongue, eroding more efficiently than the Yale Glacier at the time when the latter was a tributary of the former. The cove at the mouth of Coghill River probably has a similar hanging relationship; and College Fiord hangs above Port Wells, but how much is unknown because of the complication of glacial deposits.

Glacial Deposits. Deposits of glacial origin are not common in the College Fiord region. They include (1) ground moraine, (2) terminal moraines, (3) lateral moraines and terraces, (4) outwash gravels, and (5) deposits below sea level.

The ground moraine is everywhere thin and except for scattered erratic bowlders is entirely absent on steep slopes, but on the more gently sloping portions of the fiord walls, the deposit is thicker. The low flat strip of land on the southern side of Yale Arm has a thin veneer of glacial deposits, there being well developed morainic topography with ridges and hollows, the latter containing pools.

Since the larger glaciers end in the water, terminal moraines are few in number. Mention has already been made of the long, narrow, piedmont strip on the western side of College Fiord, built up, in part at least, by deposits from the several cascading glaciers, such as the crescentic terminal moraines south of Bryn Mawr and Wellesley Glaciers. Possibly this area also includes some deposits of the earlier, expanded Harvard Glacier. Mendenhall states[1] that in the valley of Coghill River there are "several small lakes

[1] Mendenhall, W. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, p. 273.

filling depressions back of morainal dams." The larger of these, Lake Coghill, may be held in by recessional terminal moraines of either Dartmouth or Williams Glacier.

Of the lateral moraines the one on the western side of College Point, between Harvard and Yale arms, is typical (Pl. CXXVI), consisting of till and stratified gravels, evidently built between the glacier and the mountain side. It has several terrace levels forming a morainic prow on both sides of College Point. In some places the terrace is surmounted by push moraines, in other places with ridges which may either be constructional or due to stream channeling. Most of these ridges are unquestionable push moraines, showing that the ice margin oscillated at this edge. The lowest terrace slopes southwestward from the 900 foot contour above the end of Yale Glacier to an elevation of 575 feet near the end of College Point. Gravel remnants suggesting a corresponding moraine on the eastern side of Yale Arm descend from an elevation of 1017 feet near the glacier to 200 feet about three miles to the southwest. These gravel remnants are covered by distinctive vegetation.

Above timber line, but in the zone of alder growth, on the slopes of Mt. Emerson, south of Holyoke Glacier certain shelves or benches (Fig. 43) suggest moraine terraces. In places there are three levels, with alluvial fans and snowbanks above. Between Barnard and Holyoke glaciers there is one shelf, not seen north of Barnard Glacier. Near Holyoke Glacier it is about 1500 feet above sea level. Below Mt. Emerson it slopes southward down to approximately 1200 feet, dying out opposite Coghill River, and a bench above this one extends a little farther south.

The outwash gravels near Amherst, Crescent, Williams, and Dartmouth Glaciers form the largest area of this kind of deposit in College Fiord.

Below sea level there are no definitely proved submerged moraines except the two moraine bars at the entrance to Port Wells, described in the next chapter. The 342 foot ridge or swell between Smith and Bryn Mawr Glaciers, rising 400 feet above the fiord bottom to the north, may be a submerged moraine. The 174 foot swell just west of lower Coghill River may possibly be a submerged moraine. As already indicated the shallow southern part of College Fiord seems more likely to be due to lack of excavation than to glacial deposition under water. A minor, projecting point on the western side of the fiord, near the 264 foot sounding, in this southern portion, suggests a submerged moraine bar, but the soundings are not detailed enough to determine this.

Vegetation in College Fiord. The vegetation in College Fiord furnishes important evidence concerning the episodes of former glaciation. Gilbert pointed out in 1899 some of the relationships of trees near the Yale, the Harvard, and the cascading glaciers, and stated that "a little farther south the coast is forested, and the trees climb up a few hundred feet on the moraine heaps under the hanging glaciers. They are separated from the ice, first by a broad belt of alders, and then by a barren zone. As the spruce forest in College Fiord nowhere stands close to the ice, but is separated by a barren zone, it seems fair to assume that the ice has occupied this zone so recently that the period since its shrinkage has not sufficed for reforesting; but no facts are recorded tending to show the nature of the changes immediately preceding our visit." As we have already shown, by July, 1910, most of the barren zones had been overridden.

We observed the following additional relationships of vegetation to the fiord walls

and the glaciers. In outer College Fiord near Port Wells and on the flats near Amherst Glacier, there is thick, mature spruce and hemlock forest. On the eastern side the forest extends northward nearly to Yale Glacier, ending at sea level in the cove three miles southwest of the glacier. From this point northward there is no forest at sea level, though scattered thickets of willow and alder have reached considerable size. On the eastern mountain side, however, mature spruce extends nearly up to the first tributary of Yale Glacier. It is limited above by a nearly-horizontal timberline, but its lower edge is a rather regular, sloping boundary[1] 1017 feet above sea level near the glacier and 200 feet in the cove to the southwest, practically coinciding with deposits thought to represent a faint, disconnected lateral moraine.

The western side of Yale Arm has a similar distribution of mature, thickset timber, whose growth must have taken a century or more. At the end of College Point there are scattered individuals of mature spruce at sea level, while the thick, mature forest begins at an elevation of about 575 feet, rising northeastward, with the lateral moraine already described, to 900 feet near the western edge of Yale Glacier. Above this line the forest extends up the mountain slope to the timberline; but below it are only scattered young conifers and dense thickets of alder and willow up to and beyond the end of the glacier. In these lower levels there is no thick mature evergreen forest such as extends up to the very edges of Columbia and Meares Glaciers. Some of the alders close to the front of Yale Glacier are 33 years old and the oldest willow seen had ten annual rings. From this statement it is evident that the forest conditions in Yale Arm of College Fiord are unusual, for there is a wedge of thick, mature, coniferous forest on each side, extending upward to a normal but somewhat irregular timberline, and limited below by a regular line, highest near the glacier and sloping gradually southwestward to sea level. The questions naturally arise: Is the absence of forest in the lower part of this area to be explained by unsuitability of (a) the slopes, or (b) the soil? Or is it due to a difference in the length of time during which this area has been available for tree growth? The last seems to be the case, for in the southern part of College Fiord there is thick, mature forest on all sorts of slopes and soils. The cause which until recently has kept the forest from growing in this lower belt was clearly the Yale Glacier, for an advance to or near the mouth of Yale Arm would cover the fiord walls up to about the present lower limit of the forest. The unforested area is, therefore, interpreted as a former barren zone, already thickly covered with alders and willows, some of which are 33 or more years old, while scattered young conifers have taken root where slope, soil, and adjacent seeding ground are most favorable.

This interpretation is based upon our belief that the upper belt of thick, mature forest is a remnant of a forest which formerly clothed the entire lower portion of the fiord wall, and not a forest that grew where it is while the glacier occupied the area below it. Such remnant forests exist to the west in Harriman Fiord, and we are inclined to interpret this one as of the same character, in spite of the fact that in this case we saw no push moraines with dead tree trunks at the lower and outer edges of the forest. The vegetation in Yale Arm, therefore, is interpreted as showing that the glacier extended out nearly to the end of College Point not less than 33 years ago; and, furthermore, that for a century or more it has not extended farther than that, nor higher than 900 to 1017 feet above the present terminus.

[1] Not horizontal, as Grant thought. Op. cit., p. 324.

On the western side of lower College Fiord there is continuous thick, mature, coniferous forest at sea level as far north as Barnard Glacier; but beyond that there are only scattered groves, interrupted by the barren zones of the cascading glaciers. The spruces surrounding the barren bulb of Vassar Glacier grow only near sea level, not extending up the steepened fiord walls as they do between Barnard and Wellesley Glaciers. One of the spruces at sea level, near the north side of Vassar Glacier, was 16 inches in diameter five feet above the ground and must have grown a century old or more. Between Vassar and Bryn Mawr Glaciers there are only a few spruces at sea level; but at about the hanging valley level there are two large groves. Between Bryn Mawr and Smith Glaciers there are scattered mature spruces at the hanging valley level near the former, at about the 1000-foot level near the latter, and at sea level on the south side of Smith Glacier. Between Smith and Harvard Glaciers there are scattered spruces nearly down to sea level, extending much closer to Harvard Glacier than Gilbert thought; and one which had just been pushed over by the advancing glacier in 1910 was over a foot in diameter and therefore probably nearly or quite a century old.

North of College Point, on the eastern side of College Fiord, there is thick mature forest at sea level, as far as Photo Station K (Fig. 43). North of this, to a point opposite Smith Glacier, there are large groves, also at sea level, interspersed with open spaces; and a few scattered spruces grow up to the very edge of Harvard Glacier. Throughout Harvard Arm and lower College Fiord, the open spaces are covered by dense thickets of good-sized alders and willows and, except in the narrow barren zones of the cascading glaciers, there is no considerable area with as sparse growth as on the eastern side of Yale Arm.

Thus Harvard Arm has much more forest than Yale Arm, and the eastern side is more thickly forested than the western where the cascading glaciers have broken it up. The conditions in Harvard Arm show clearly that Harvard Glacier has not recently extended southward beyond Pt. K., where the thick forest ends, and that for at least a century no advance has gone as far as the one which was in progress in 1910, contrasting, therefore, with the recent behavior of Yale Glacier.

Reason for Advances of Glaciers

It is not perfectly clear why the glaciers of Unakwik Inlet and College Fiord are oscillating, some advancing and some retreating. The fundamental cause is, of course, involved with snow supply and with rate of melting as determined by temperature. Climatic records are so incomplete that we cannot settle this question and it is quite possible that, as in Yakutat Bay, the snow supply of any glacier, or group of ice tongues, might have been increased by avalanching during earthquakes, rather than by increase of precipitation. These glaciers of northern Prince William Sound lie entirely outside the zone affected by great avalanching during the Yakutat Bay earthquakes of September, 1899; but there have been a number of other severe, recent earthquakes in Alaska, whose origins were in or near the region of these advancing glaciers. One such, in October, 1900, was of considerable intensity.

The Yakutat Bay glaciers which have so far advanced as a result of the 1899 earthquakes are of variable sizes and have responded in order of size, the smaller ones first.

The advancing glaciers in Prince William Sound are also of variable sizes; but here the largest, Columbia and Valdez, seem to have begun to advance in 1907–08, long before the smaller College Fiord and Unakwik Inlet ice tongues. Here, nine or more glaciers, the large Harvard, Yale, and Meares Glaciers, and the smaller, Wellesley, Vassar, Bryn Mawr, Smith, Radcliffe, Barnard, and possibly other small ones, all began to advance together in 1909–10. This absence of relation between size and period of advance, and the lack of spasmodic transformation to activity with rapid return to stagnation, observed in Yakutat Bay, suggests that the advances now in progress in this part of Prince William Sound are probably climatic.

The following table [1] shows the precipitation record from the nearest stations. It also suggests that the glacier advances in College Fiord and Unakwik Inlet are climatic. The data is given both in snowfall and in total precipitation (rainfall plus melted snow), and is arranged by years from July 31 to July 31, so that the snowfall of the winters of two calendar years appears together, that is by snow years.

Table Showing Precipitation from (1) Fort Liscum Near Valdez, (2) Sunrise on Turnagain Arm, (3) Kenai on Cook Inlet, All at Sea-Level, and (4) Copper Center, 1000 Feet Above Sea-Level

Year	Total Precipitation, in inches	Snowfall, in inches			
	Fort Liscum	Fort Liscum	Sunrise	Kenai	Copper Center
1901—02	78.67	334.4	——	32.3	——
1902—03	96.26	671.1	——	109.1	36.5
1903—04	57.38	255.9	89.4	92.6	47.1
1904—05	69.58	187.8	123.4	40.1	25.5
1905—06	72.62	338.5	112.1	43.7	64.1
1906—07	72.31	299.0	83.8	47.1	28.3
1907—08	91.35	447.2	158.9	——	31.5
1908—09	73.52	338.7	85.0	——	27.0
1909—10	70.29	458.5	132.2	——	33.0
Mean	75.77	370.1	112.1	60.8	36.6

It might be thought that the increase in snowfall at Fort Liscum (a) from 334 inches in 1901–1902 to 671 inches in 1902–1903, an augmentation of 337 inches or 28 feet, or (b) from 299 inches in 1906–1907 to 447 inches in 1907–1908, an increase of 148 inches or 12½ feet, or (c) from 338 inches in 1908–1909 to 458 inches in 1909–1910, an increase of 120 inches or 10 feet, would be quite adequate to account for the advance of the eight glaciers in College Fiord. To this there are two possible objections.

First, Fort Liscum, the nearest of the U. S. Weather Bureau observatories, is nearly 50 miles from College Fiord (Pl. XCIII); and it is by no means certain that the climatic oscillations are similar and synchronous. Unfortunately the other places where precipitation records are kept, to the north and southwest of College Fiord, are in regions of dissimilar climate and the records are incomplete, though the table just quoted shows

[1] Martin, Lawrence, Some Features of Glaciers and Glaciation in the College Fiord, Prince William Sound, Alaska, Zeitschrift für Gletscherkunde, Band VII, Heft 5, 1913, pp. 28–31.

that the snowfall (a) at Kenai, 130 miles southwest of College Fiord increased from 32 inches in 1901–1902 to 109 inches in 1902–1903, corresponding to the great increase at Fort Liscum in 1902–1903, and (b) the snowfall at Sunrise, 65 miles southwest of College Fiord, increased from 83 inches in 1906–1907 to 158 inches in 1907–1908, corresponding to the increase at Fort Liscum in the latter year. At Chickaloon, 950 feet above sea level in the Matanuska valley north of the Chugach Mountains and only 45 miles northwest of College Fiord, we have snowfall records during only two winters. These show 95½ inches in 1907–1908 (October 25 to March 31) and only 75½ inches during the corresponding months of 1910–1911. At Copper Center, 90 miles northeast of College Fiord, the increased snowfall appears for 1907–1908, but there was not as much in 1902–1903 as in the following year, and the greatest snowfall was in 1905–1906. Ten out of fifteen other stations in distant parts of Alaska, however, show a great increase of snowfall in 1902–1903 and 1907–1908.

Secondly, the two glaciers extending down to sea level close to Fort Liscum show no adequate response to the increase in snowfall in 1907–1908. Valdez Glacier is only 8 miles north and Shoup Glacier only 10 miles northwest of Fort Liscum. Following the doubling of snowfall in 1902–1903, however, the Valdez Glacier had a slight advance between 1905 and 1908; but the adjacent Shoup Glacier had no such advance, and Valdez Glacier moved forward only 250 to 300 feet and then commenced to recede again, as it has done ever since and as we know from annual observations that it has been doing ever since 1898. Columbia Glacier, midway between College Fiord and Fort Liscum, began to advance in 1908 and continued slow forward motion at least until 1911.

All this suggests that the tremendous annual snowfall of 187 to 671 inches (15½ to 56 feet) at Fort Liscum, and the smaller amounts at Sunrise, Kenai, and Copper Center, may not be representative of the snowfall in the névé fields of the adjacent glaciers, where the precipitation is probably even greater, but does not necessarily fluctuate in exactly the same way.

On these accounts, though it is probably, it is not absolutely safe to ascribe the forward movement of the Meares Glacier in Unakwik Inlet and the eight glaciers in College Fiord in 1910 to the increased precipitation recorded at Fort Liscum during the winters of 1902–1903 or 1907–1908. The temperature records are open to the same limitations.

The only satisfactory attitude to take at present is to ascribe this great renewal of activity of certain of these glaciers either (a) to climatic variations or (b) to earthquake avalanching, and then wait to see whether other adjacent glaciers, now retreating, such as the Amherst in College Fiord and the Barry in Harriman Fiord, also advance, and whether those now advancing continue to advance for several years as the Columbia Glacier has done, suggesting a climatic explanation, or whether they develop the characteristics of the earthquake-stimulated Yakutat Bay glaciers, having spasmodic advances which are of short duration. It would be most desirable if some one could revisit the Unakwik Inlet and College Fiord glaciers in a year or two to gather information on this important question.

The initiation of what seems to be a general advance of the glaciers of western Prince William Sound is a matter of no little importance. Fourteen ice tongues began to advance in 1909–1910—eight of them in College Fiord—and a fifteenth had commenced to move forward in 1908. The importance lies (a) in distinguishing whether this advance is climatic or due to earthquake avalanching; (b) in determining whether the advance is

a minor interruption of the previous slow retreat, as the increase in snowfall for only one year at a time would suggest; and (c) in affording an opportunity, in case the advance is climatic and is not a small local affair, of seeing how long after an increase in snowfall the general advance begins and for what period of years it continues. In the latter case the advance might go on for some considerable part of the period of 35 to 50 years observed in many glaciers of the Alps and other mountain regions.[1] It is especially interesting as possibly yielding future data in connection with the 35 year periods described by Brückner.[2] It would be most desirable if some one could revisit the Unakwik Inlet and College Fiord glaciers and the others in Harriman Fiord and the remainder of western Prince William Sound after a few years to gather information on this important question.

[1] Forel, A., Jahrbuch des Schw. Alp. Verein, Annual volumes beginning in 1881.

Rabot, C., Revue de Glaciologie, Ann. du Club alpin Français, Vol. XXVIII, Paris, 1902; *Ibid.*, No. 2, Op. cit., Vol. XXIX, Paris, 1903; *Ibid.*, No. 3, Mémoires de la Société Fribourgeoise des Sciences naturelles, Vol. V, Fribourg, 1909.

Finsterwalder, S., Comptes rendus du Congrès géologique international, de Vienne, 1903, pp. 161-169.

Reid, H. F., Variations of Glaciers, Jour. Geol., Vol. III, 1895, pp. 278-288; and annually since then in the Journal of Geology. See also the annual volumes of the Zeitschrift für Gletscherkunde.

Brückner, Eduard, La Commission internationale des Glaciers au Congrès géologique international, Stockholm, août, 1910, Zeitschrift für Gletscherkunde, Band V 1911, pp. 161-176.

[2] Brückner, Eduard, Klima-Schwankungen seit 1700, Geog. Abhand. von Dr. Penck, Band IV, Wien, 1890, pp. 183-193.

CHAPTER XVI

GLACIERS OF HARRIMAN FIORD AND PORT WELLS

Harriman Fiord

General Description. Harriman Fiord is the northwestern arm of Port Wells, the northeastern branch being College Fiord. It has the form of a bent arm. The lower part of this fiord from Port Wells to the elbow is called Barry Arm. It extends northwestward from Pt. Pakenham for a distance of about 8 miles, Barry Glacier being at the elbow end (Map 8).

Harriman Fiord proper extends westward from the elbow of Pt. Doran at right angles to Barry Arm and then turns southwest, having a total length of over 12 miles, and terminating at the tidal front of Harriman Glacier. About half way down the western side is a tributary fiord 2½ miles long, terminated at the western end by Surprise Glacier. On the southern side of this inlet is Cataract Glacier, which descends the fiord wall to tidewater. About half way between Surprise and Barry Glaciers is the fifth tidal ice tongue of Harriman Fiord, Serpentine Glacier, which ends in a small cove. The other ice tongues of the fiord, none of which reach the sea, are Toboggan, Baker, Detached, Roaring, Dirty, Wedge, and a number of smaller ones. The snow line is at an elevation of 2000 to 2500 feet (Pl. CXVII, B), and above it the proportions of snow covering and bare rock vary with the slopes.

Barry Arm and Harriman Fiord are from 1 to 3 miles wide and are bordered by steep fiord walls, rising from 3000 to 4000 feet within a mile of the fiord and with many peaks attaining greater heights, including Mt. Gilbert, 10,194 feet high, the highest peak in the western Chugach Mountains (Pl. CXXVII), at a distance of 5½ miles from the fiord, Mt. Gannett,[1] 9240 feet high (Pl. CXXVIII), 5 miles from the fiord, and Mt. Muir, 8207 feet, 2 miles from the fiord. Harriman Fiord lies in the very heart of the Chugach Mountains and the northern slopes of Mts. Gilbert, Muir, and Gannett supply ice for the Knik Glacier which flows northward to the Matanuska Valley and the head of Cook Inlet. A high glacier between Mt. Gilbert and Mt. Gannett, visible from Port Wells, is probably a part of the north-flowing glacier system.

Previous Studies. The Barry Arm portion of Harriman Fiord was shown roughly upon maps by Vancouver in 1794,[2] Applegate in 1887,[3] and Glenn,[4] Castner [5] and Men-

[1] Named in 1910 for Henry Gannett, President of the National Geographic Society, who was a member of the Harriman Expedition in 1899 and who first mapped the glaciers of this fiord.

[2] Vancouver, Capt. George, A Voyage of Discovery to the North Pacific Ocean and Round the World, Vol. V, 1801, p. 312; map also in Davidson (see below), Pl. V.

[3] Applegate S., map in Davidson's Glaciers of Alaska that are Shown on Russian charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc. Pacific, Vol. 3, 1904, Pl. XI.

[4] Glenn, E. F., War Dept., Adj. Gen. Office, No. XXV, 1899, pp. 19, 21, and map (in pocket).

[5] Castner, J. C., *Ibid.*, p. 191.

denhall[1] in 1898, the Barry Glacier being briefly mentioned in the later accounts. In 1794, 1887 and 1898 Barry Glacier extended nearly across to Point Doran, so that, from near Pt. Pakenham, at the northern end of Port Wells, Barry Arm seemed to terminate there.

It was not until 1899 that the existence of the southwestern portion of the fiord was discovered by white men, though Grant states that native seal hunters had previously entered it. In that year, the Harriman Expedition, going close to the front of Barry Glacier, discovered the inlet to the southwest which they entered and explored. The Harriman, Cataract, Surprise, Serpentine, and smaller glaciers were described for the first time by various members of the party,[2] including G. K. Gilbert.[3] A map was made by Gannett[4] (Pl. CXXIX, A), and the glaciation of Harriman Fiord and Port Wells was studied.

In 1905, 1908 and 1909 Harriman Fiord was visited by Grant,[5] Paige and Higgins, who remapped the glaciers, taking many photographs, describing the great retreat of Barry Glacier from 1899 to 1909, and giving many new facts about the glaciers and glaciation of the region.

The National Geographic Society's expedition of 1910 spent July 25th to August 2d in Harriman Fiord and Port Wells, studying the glaciers and the former glaciation,[6] making special maps of a number of the ice tongues, and soundings throughout Harriman Fiord and Port Wells.

Barry Glacier. Barry Glacier, heading in snowfields on Mt. Gannett and unnamed peaks 8000 to 9000 feet high, west of the cascading glaciers of College Fiord, has a known length of over 9 miles, and is about a mile wide in its lower valley. Close to the terminus it is joined by the Cascade and Coxe Glaciers, cascading tributaries each half a mile wide, which increase the width of the main glacier to a mile and three eighths at the terminus. The glacier ends in a wide ice precipice between 200 and 300 feet high, discharging small icebergs. The slope of the main glacier surface is 350 to 400 feet to the mile and it is severely crevassed from side to side.

The Coxe and Cascade tributaries, severely-crevassed, cascading glaciers, slope 2000 to 3000 feet to the mile. The former, heading in cirques near Mt. Coville[7] and Mt. Emerson, has no medial moraines. There is a lateral moraine on the west side but none on the east. This western lateral moraine coalesces with a lateral and east medial of

[1] Mendenhall, W. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, p. 325 and Map 16.

[2] Gannett, Henry, Nat. Geog. Mag., Vol. X, 1899, pp. 510–511; Bull. Amer. Geog. Soc., Vol. XXXI, 1899, pp. 346–347, 354–355; Harriman Alaska Expedition, Vol. II, 1901, p. 263; also Muir, John, *Ibid.*, Vol. I, 1901, pp. 132–133; also Burroughs, John, *Ibid.*, Vol. I, 1901, pp. 71–74. Many of the photographs most useful in connection with the subsequent retreat of the glaciers are by C. Hart Merriam.

[3] Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, pp. 89–97, 174, 176.

[4] *Ibid.* Pl. XIII, facing p. 80.

[5] Tidewater Glaciers of Prince William Sound and Kenai Peninsula, Bull. U. S. Geol. Survey (in preparation); in Reid's Variations of Glaciers, Journ. Geol., Vol. XIV, 1906, p. 407; *Ibid.*, Vol. XIV, 1908, p. 671; also maps in Bull. 284, U. S. Geol. Survey, 1906, Fig. 4, p. 79; *Ibid.*, Bull. 379, 1909, Pl. IV facing p. 88; *Ibid.*, Bull. 443, 1910, Pl. I, facing p. 10, and Pl. II, in pocket; Glaciers of Port Wells, Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLIII, 1911, pp. 327–338.

[6] Martin, Lawrence, Nat. Geog. Mag., Vol. XXII, 1911, pp. 550–551, 553–560; Gletscheruntersuchungen längs der Küste von Alaska, Petermann's Geog., Mitteilungen, Jahrgang 1912, Septemberheft, pp. 147–149; Collier's Weekly, Vol. XLVII, No. 17, 1911, p. 20.

[7] Named in 1910 for Frederick V. Coville who visited this region in 1899 with the Harriman Expedition.

Meares Glacier in 1910
Photographed from Station D (Fig. 37). Shows advancing northern margin.

Plate CXIV

A. General View of Harvard Glacier
Showing portion of Chugach Mountains which it drains. Photograph from Station N (Map 7), July 22, 1910.

B. Inclined Ends of Medial Moraines of Harvard Glacier
Detailed view of part of ice cliff. Photograph, July 17, 1910.

Plate CXV

Western Margin of Harvard Glacier
Advancing into forest Photograph, July 21, 1910.

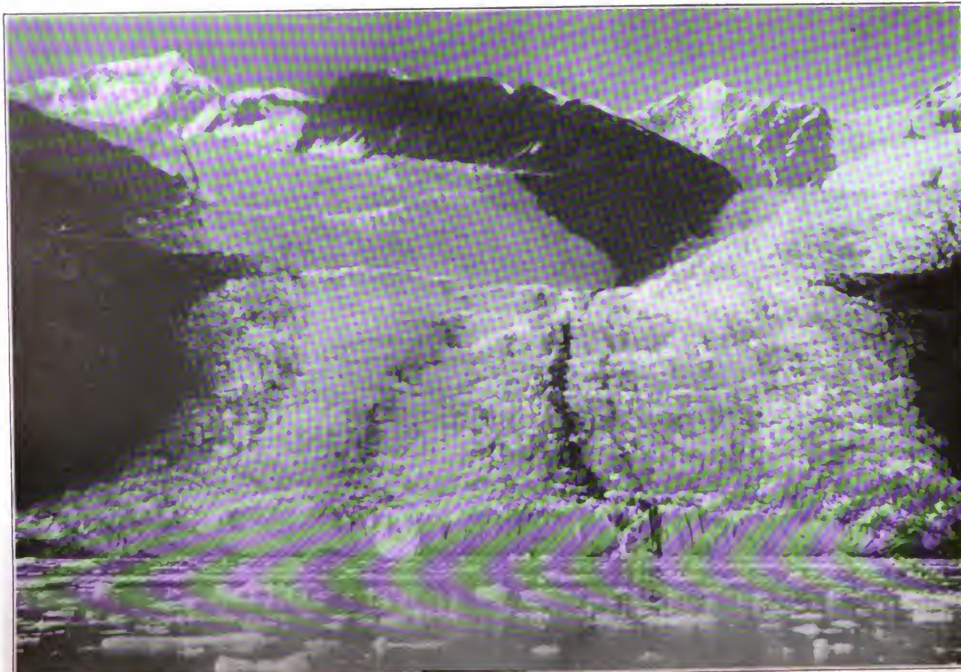
PLATE CXVI

GLACIER ADVANCING INTO ALDER THICKET
On northern margin of Smith Glacier. Photograph, July 22, 1910.

A. Flattened Piedmont Terminus of Smith Glacier
Where it entered the sea in 1910. Photograph, July, 1910, from north side.

B. General View of the Less Rugged Topography in Harriman Fiord Region
Showing Barry Glacier and cascading tributary. Snow line of August, 1910, at elevation of between 2000 and 2500 feet. Mt. Curtis on right, Mt. Coville above right-hand margin of Barry Glacier. Photograph, July 26, 1910.

Plate CXVIII



Bryn Mawr Glacier

Showing vertical terminal ice cliff and relationship of medial moraines in tributaries. Photograph by Grant and Paige, 1905, from Photo Station O.

PLATE CXIX

PROFILE OF THE FLATTENED PIEDMONT PORTION OF VASSAR GLACIER
Covered with ablation moraine. View from north, July 21, 1910.

PLATE CXX

WELLESLEY GLACIER

Showing snowfields and typical cascading glacier. Part of barren zone not yet covered. Photograph from Station R, July 22, 1910.

Plate CXXI

Yale Glacier from College Point
Ink lines show 750 foot advance of eastern edge between 1899 and 1910, chiefly in the last year. Castner Glacier and Mt. Castner on right. Photograph, 1899, by Curtis of Harriman Expedition, from Photo Station P.

PLATE CXXII

YALE GLACIER FROM PHOTOGRAPH STATION I
The best picture to use for determining future advance or retreat, July 15, 1910.

EASTERN SIDE OF YALE GLACIER
Showing snow on sea ice, and ledges beneath the glacier, as in 1910. Photograph by W. C. Mendenhall, April, 1898.

PLATE CXXIV

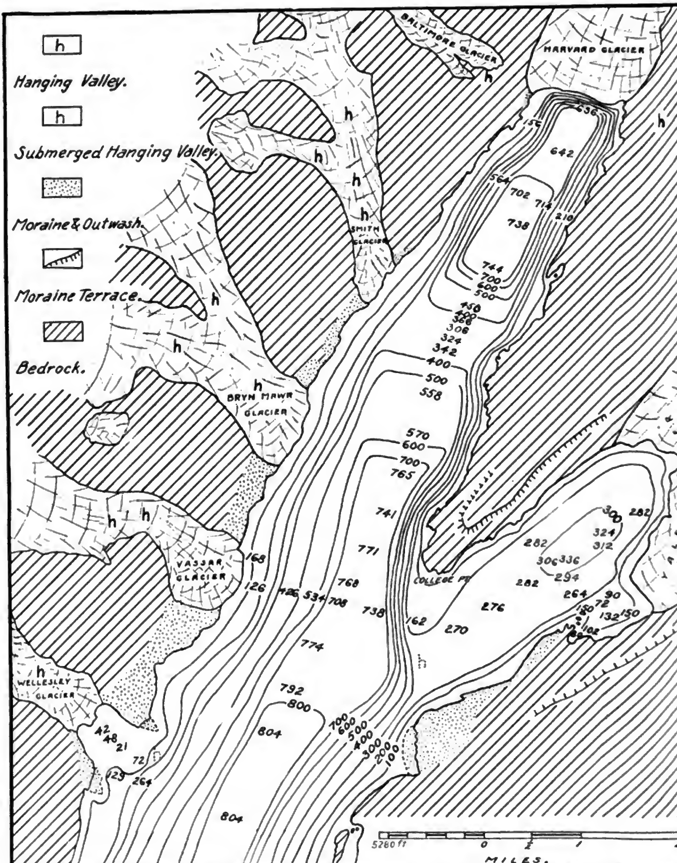
A. General View of Yale Glacier
Photograph by C. Hart Merriam, 1899.

B. Amherst and Crescent Glaciers
From southern part of College Fiord. Photograph, July 25, 1910.

PLATE CXXV

HANGING VALLEY IN COLLEGE FIORD

With Wellesley Glacier cascading down over its lip to tidewater, 1700 or 1800 feet below. From Photo Station O, July 22, 1910.



SUBMARINE TOPOGRAPHY OF NORTHERN COLLEGE FIORD

Soundings in feet, contour interval 100 feet. Some features of glacial erosion and deposition above sea level also shown.

PLATE CXXVII

ROUGHER TOPOGRAPHY OF CHUGACH MOUNTAINS IN HARRIMAN FIORD REGION

Mount Gilbert, 10,194 feet. Shows large proportion of bare rock above snow level in region of steep slopes at heads of cirques and on oversteepened walls of valleys containing active glaciers. Photograph from Point Pakenham at head of Port Wells, August, 1910.

PLATE CXXVIII

BARRY GLACIER AND MOUNT GANNETT, 9200 FEET

Shows large proportion of snow-covered area above the snow line in region of moderate slopes. Photograph from Station A, July, 1910.

the main glacier, forming a weak medial moraine which terminates near the eastern side of the glacier terminus. Near the western margin of Barry Glacier is an unusually strong, black medial moraine, separated from the valley wall by a strip of clean ice and by a strong black lateral moraine.

Cascade Glacier (Pl. CXXVIII), heading in 6000 to 8000 foot cirques on Mt. Gannett, descends over the lip of its hanging valley at a level of about 3000 feet with a lower step at 2100 feet. The western side of Cascade Glacier has a dark lateral moraine which becomes the western lateral of Barry Glacier below their confluence. The eastern side has none.

Just above the entrance of Cascade Glacier the edge of Barry Glacier is forced up on a rock surface over which its western lateral moraine rides with a fine ribboned effect; but below this point the broad black moraine all but disappears, showing on the terminal ice cliff only as a faint, narrow band, slightly inclined. This disappearance of the moraine seems to be due to sliding down of fragments in the more severely crevassed lower end of the Barry Glacier where it is joined by the steeply-descending Cascade Glacier.

Below their confluence the Cascade Glacier is compressed to half width by the stronger Barry Glacier, as the medial moraines prove. A broad medial moraine which remains nearly vertical at the terminal ice cliff forms a striking contrast to the inclined ends of the Harvard Glacier moraines. The Barry and Harvard and the Cascade and Radcliffe Glaciers have similar slopes; but the Radcliffe cascading glacier retains nearly full width at the ice front, compressing the main Harvard Glacier to half width, while in Harriman Fiord these relationships are reversed.

There are several small glaciers near the Barry Glacier. The group between Mt. Coville and Mt. Curtis[1] terminate 3000 feet above sea level. Two in cirques and one in a hanging valley between Barry and Serpentine Glaciers terminate 2350 to 4000 feet above sea level.

Recession. The position of Barry Glacier in 1794 was apparently about the same as in 1887, judging by Vancouver's map. We have very little information about the condition of the glacier at this early date, Lieutenant Whidbey speaking of Barry Arm as "a bay, about a league and a half deep toward the N. N. W., in which were seen several shoals and much ice."

Applegate made no description of Barry Glacier in 1887. He went no nearer to it than Pt. Pakenham, marking the termination in about the same position as shown on the map of 1898, and showing no indication of the existence of Harriman Fiord.

Mendenhall relates that in 1898 Barry Glacier was more extensive than either Harvard or Yale Glaciers. Glenn, who in the same year named it for Col. Thomas H. Barry, said that it was "one of the most formidable as well as the most interesting" of the glaciers that they saw. Coming from it they saw "immense icebergs that had evidently broken off from the glacier. Many of these were from ten to twenty times as large as our boat."[2] When they were near Yale Glacier, alternating two or three times an hour with the roar of snowslides, a "noise caused by the falling of the immense ice floes from Barry Glacier, could be heard like the rumbling of distant thunder and which

[1] Named in 1910 for E. C. Curtis who took many excellent photographs of this and other Prince William Sound glaciers, at the time of his visit with the Harriman Expedition in 1899.

[2] The *Salmo*, a small steamer.

seemed to shake the mountains on either side of us. We must have been at least 15 miles from Barry Glacier, and yet we could distinctly hear the icebergs falling from it into the sea." In May, 1898, Castner went up Barry Arm "to within a half mile of the sea end of the great glacier. Photographs were taken and the interesting manufacture of icebergs watched. The latter consisted of a breaking off and tumbling into the sea of tons of blue ice from the face of the glacier, accompanied by the roar of a Niagara, as the berg started on its ocean voyage, eventually to melt and become a part of the tides which now carried it away."

This description by Castner, with the mention of the great size of icebergs and the noises heard by Glenn from as far as Yale Glacier, a phenomenon not observed in 1910, are quoted because they suggest some very great activity in 1898. Field evidence, observed in 1910, in connection with a study of the distribution of the vegetation, independently fixed 1898 as the date of the last great activity, which seems to have ceased before the visit of the Harriman Expedition the following year.

It is possible that Barry Glacier barred Harriman Fiord in 1898, perhaps all but touching Pt. Doran. But the absence of a barren zone on Pt. Doran proves that the glacier did not actually rest upon this point. This is probably because of the extremely strong tides which were observed here even in 1910 after Barry Glacier had retreated 3½ miles. In 1899 the glacier front stood about ¾ of a mile from Pt. Doran, as Gannett's map shows. The narrative and photographs by members of the Harriman Expedition indicate that the discharge of large icebergs, and the unusual noises observed by Castner and Glenn a year before, had ceased. Gilbert's description of the 1899 conditions is as follows:

"Barry Glacier, at the entrance to the fiord, approaches from the north-northeast. Its low grade indicates a distant source . . . it impressed the beholder as one of the largest ice rivers of Port Wells. Its peculiar relation to the fiord causes it to be swept by the passing tide and prevents the accumulation of icebergs about its front, but the same relation exposes it to exceptionally rapid melting by the sea, and the conflict of ice current with tidal current must be active. The forward flow of the ice tends to narrow the strait, and this constriction, by increasing the speed of the tide, enhances the melting power of the water. The fact that the glacier was able to occupy two thirds of the width of the fiord indicates that its forward movement was strong.

"Its moraines were of small relative importance, but a belt along the western margin was darkened by drift and there were two medials. One of the latter, exhibited in section in the face of the cliff, was seen to be the surface outcrop of a sheet of drift-charged ice which extended obliquely downward, passing under the western portion of the stream.

"The cliff was further diversified by a number of caves at the water's edge, supposed to be the mouths of englacial streams.

"Connected with the eastern edge of the ice was a long, narrow tongue attached to the shore (Pl. CXXX, A), evidently a remnant left by the glacier at some very recent date when its front was more extensive. As this strip was not protected by gravel, it must have been wasting rapidly, and the period of its separation may have been only a few months. I was in doubt whether to ascribe it to a progressive shrinkage of the glacier or to seasonal variation. On the same coast the forest did not approach the glacier closely at the water line, but passed above it, leaving a barren zone

several hundred feet broad. The common boundary of the barren zone and forest was so well defined as to indicate that it represented a former limit of the ice, but there were no overturned trees. If the forest ever occupied the barren zone, and was there destroyed by an advance of the glacier, the occurrence was so long ago that the overturned tree trunks had disappeared through decay. The portions of the forest nearest the ice included no trees of large size, but as there were many standing dead trunks it is probable that the growth was mature and that the small size of the trees indicated merely conditions unfavorable to luxuriant growth. Translating these facts into terms of glacial history, it seems probable that the Barry had been, at sometime within the century, somewhat larger than when we saw it, but that it has not for a series of centuries exceeded the limit marked out by the neighboring forest. If any change had occurred within the last year or two it was of diminution.

"The opposite wall of the fiord is forested down to the water's edge, and it is thus shown that no recent advance of the glacier has carried it completely across the channel."

Fig. 44. Front of Barry Glacier.

Showing recession from 1899 to 1909 (After Grant and Higgins).

By 1905 Grant reports that "the extreme front of Barry Glacier . . . has retreated at least a mile [1] since 1899. The long front which projected into Doran Strait for nearly two miles has entirely disappeared, and the front is now nearly straight across. The little tongue of ice which lay along the east side of the glacier has also disappeared." A photograph from Pt. Doran by Grant and Paige in 1905 (Pl. CXXX, B) shows this retreat graphically in comparison with a photograph by Merriam, from nearly the same site, showing enormous changes both on the southwest and the southeast. Barry Glacier was revisited by Grant in 1908 and "its front was found to have retreated on the east side about a fourth of a mile, and more than this on the west side." [2] Grant's 1909 photographs (Pl. CXXXI, A) show that a year later the retreat of the western edge had increased about 880 feet, and he has stated that the middle of the glacier had retreated half a mile. At this time Higgins made the map of Barry Glacier reproduced in Fig. 44, also showing graphically a decade of this retreat.

In July, 1910, the National Geographic Society's expedition found that the amount of retreat of Barry Glacier during the previous year was about 1600 feet on the eastern side and 500 feet on the western. There were disconnected ice masses above and in front of the glacier, in the area of fiord wall exposed by melting during that year.

At the terminus of the glacier in 1910, the vertical thinning by ablation during the $3\frac{1}{2}$ mile retreat from 1898 to 1910, amounted to 900 feet on the western and 1000 feet on the eastern side. Between the Cascade and Barry Glaciers, the ice surface was lowered 650 feet, and between Coxe and Barry Glaciers 700 feet. The curving medial moraines

[1] Given in a later publication as 1.2 miles.

[2] About 2200 feet on the western side, as measured on a photograph. Grant and Higgins state that the retreat in this period was four-tenths of a mile along the axis of the glacier.

that formerly existed on the ice between the 1899 and 1910 terminus, seen in the 1899 photographs, were of course all removed by the shortening of the glacier.

In 1910 the terminus of Barry Glacier was, therefore, over $3\frac{1}{2}$ miles farther north than in 1899 (Fig. 44). The stages of this retreat are shown in photographs taken in 1899, 1905, 1909, and 1910 (Pls. CXXX and CXXXI) and in the maps (Pl. CXXIX). The amount of ice lost during this retreat amounts to over a cubic mile.

The behavior of Barry Glacier throughout the period of observation may, accordingly, be summarized as follows:

Date of Observation	*Change*	*Amount*	*Observer*
Before June 8, 1794	Unknown		Whidbey
June 8, 1794, to June, 1887	Little known change		Applegate
June, 1887, to May 6, 1898	Advance in 1898		Castner
May 6, 1898, to June 26, 1899	Retreat	$\frac{1}{2}$ to $\frac{3}{4}$ mile	Gannett
June 26, 1899, to Aug. 20, 1905	Retreat continued	1 $\frac{2}{10}$ mile	Grant and Paige
Aug. 20, 1905, to Aug. 11, 1908	Retreat continued	$\frac{4}{10}$ mile	Grant
Aug. 11, 1908, to June 29, 1909	Retreat continued	Half a mile	Grant and Higgins
June 29, 1909, to July 25, 1910	Retreat continued	500–1600 feet	Martin

Detached Ice. A conspicuous feature associated with the recent extension of Barry Glacier is a detached ice mass $\frac{1}{4}$ mile wide and over $\frac{1}{8}$ mile long, which in 1910 was two miles from the ice front. It had evidently lingered because of the protection of the broad lateral moraine on the Barry Glacier shown in an 1899 photograph. In 1910 it was a slumping, irregular, ablation moraine covered ice mass, extending up the hillside to a height of 240 to 285 feet, and on its southern and northeastern sides having a steep margin of glistening black ice, thickly-set with stones. Besides this isolated ice mass, there was to the south and west an additional adjacent area where slumping was going on, young shrubs and moss were being overturned, and much water was emerging, furnishing evidence of the presence of deeply-buried ice.

There were also subordinate, minute, detached masses of ice nearer the front of Barry Glacier, those on the western side being close to the ice front and covered with ablation moraine. In most places the western fiord wall is too steep to retain ice masses of this sort.

Barren Zone. A second feature proving recent retreat is an extensive barren zone, the margin of which was already visible in 1899 when Gilbert discussed it briefly. Its increased area from 1899 to 1909 was described by Grant and Higgins, and indicated on a map (Fig. 44). It is shown in still greater detail in Pls. CXL and CXXXIX. The upper edge of this barren zone reaches a height of 950 to 1100 feet above sea

level. It includes a strip of fiord wall from an eighth to a quarter of a mile wide in the area occupied by the Barry Glacier in 1898, widening at the southwest, opposite Pt. Doran, in a morainic flat ¾ of a mile wide and nearly 2 miles long.

Much of the barren zone is made up of smooth glaciated rock surfaces, with scattered erratics and an absence of all vegetation except a very few shrubs, none of them more than 11 years old. At the upper margin of the barren zone, on each side of Barry Arm, there is a well-marked push moraine, at the edge of the dense mature forest into which the glacier advanced during the last maximum.

Moraines. There are well developed terminal and lateral moraines at the 1898 terminus of Barry Glacier. Part of the terminal moraine is below sea level, but a larger proportion than in most other tidal glaciers of this region is on the land and this terminal moraine, which forms the large triangular flat northwest of Point Doran, covers an area of 1¼ miles long and over ½ mile wide, and projects southward into the fiord for 3000 feet, narrowing this portion of Harriman Fiord to less than half its normal width. At low tide it is continued southward 1800 feet by a submerged continuation, and southeast of this across to Pt. Doran the fiord is nearly filled. Above this submerged moraine three shoals rise out of the water at low tide, their surfaces made of till and glacial bowlders upon which many icebergs strand (Pl. CXL).

The dry-land portion of this terminal moraine has numerous kettles and ridges and there are three lines of ponds. It constitutes a low morainic surface with an area of outwash gravels at the northern edge, near the mountain wall. At the western edge these grade into beach deposits of sand and mud which fringe the whole southwestern margin of the moraine. An abandoned stream channel, with coarse waterworn gravels, leads westward from the edge of the isolated glacier remnant, where a large glacial stream evidently emerged from the ice in 1898 or 1899. The eastern border of the moraine grades into slumping material, still being added in 1910 by the melting of the large stagnant ice block. East of the ice block the terminal moraine rests upon the smooth rock surface of the barren zone, much of which slopes so steeply that the débris left by the glacier has slid down, forming an accumulation at the mountain base. There are only scattered shrubs on this terminal moraine, the oldest having 11 annual rings.

A lateral moraine ridge extends up the mountain side from the terminal moraine area rising northward toward Barry Glacier. It slopes upward at the rate of 1650 feet to the mile for the first third of a mile, and 932 feet to the mile for the next five eighths of a mile. Beyond this we have not traced the lateral moraine; but the edge of the barren zone first becomes flat for a mile and a half, then descends 300 feet, and then ascends 150 feet. This proves that the slope of Barry Glacier was much flatter in its expanded condition in 1898 than in 1910 (350 to 400 feet to the mile), but that it sloped much more steeply at the terminus than now.

This lateral moraine is generally a single ridge (Pl. CXXXII, A), rising in places 12 to 15 feet, in other places 25 to 30 feet above the surrounding surface. It seems to be a push moraine like those at Columbia Glacier, and is so narrow for most of its length that it is often difficult to walk upon its somewhat sinuous crest. Rarely it broadens to 50 feet, though generally retaining a narrow crest and very steep lateral slopes. Less than a mile from its beginning it terminates against a precipitous rock cliff upon which no moraine could stand. This is just south of a prominent waterfall (Pl. CXL). Much of the constituent material is till and rounded gravel but there are also many

large striated bowlders and some buried trunks of trees, some of them 15 to 32 inches in diameter. Outside this moraine there are occasional peat rolls and, rarely, duplicate parallel push moraines.

Upon this western lateral moraine are only scattered shrubs including eleven year old alders and younger spruces; but extending down to the very edge of the moraine is a thick, mature, coniferous forest with closely-spaced trees a foot or two in diameter, deep moss and peat, and a luxuriant growth of grasses, forming a striking contrast to the barren zone below. Although the overwhelming majority of trees outside the barren zone are alive, there is an occasional dead tree standing erect, and there are many which were overturned, but not overridden, by the ice edge which formed the lateral moraine. Some of these trees have fallen outward into the forest; some fell toward the ice and rest upon the moraine; and in places there is an inextricable tangle of fallen trunks. The conditions in 1910 vividly recalled those seen a few weeks before along the margins of Columbia Glacier which was then advancing into the forest, as Barry Glacier must have been doing during and just before 1898.

Along the eastern shore of Barry Arm, there is a similar lateral moraine at the edge of the barren zone. It is usually from 6 to 10 feet high, but rises higher where it crosses the mouths of gullies. It extends up to the very edge of thick mature forest, and contains till, gravel, bowlders and wood, as in the case across the fiord. In places this margin has windrows of dead tree trunks, some of them 32 to 40 inches in diameter. On the moraine are young alders, willows, and spruces. We counted the annual rings in one spruce 3 inches in diameter, the largest seen growing on the moraine, and found it to be 11 years old. Outside the moraine were some dead trees still standing erect, and a number of spruces and hemlocks whose tops had been broken off 15 to 25 feet above the ground evidently by ice splinters projecting beyond the lateral moraine, as observed in progress at Columbia Glacier. The broken tree tops lay at the foot of the trunks, which were sometimes inclined and sometimes erect. In several cases the broken tree had sprouted a new top, two or three inches in diameter and always growing erect even on inclined trees. Outside the moraine the forest had thick moss and gave every indication of long, uninterrupted growth. A tree, newly sawed off, which was 22½ inches in diameter had 225 annual rings, suggesting that the fallen trunks which measure 32 to 40 inches in diameter were not less than two centuries and perhaps three or four centuries old. This furnishes a basis for computing the length of time that has elapsed since there has been a glacial advance as extensive as that ending in 1898.

Up to an elevation of 1000 feet this eastern lateral moraine slopes at an average rate of 450 feet to the mile, then flattens to horizontality; ascending again north of Coxe Glacier. The Harriman Expedition photographs and Gilbert's description prove that since the maximum of this advance there had been sufficient melting so that by 1899 the ice had shrunk away from the edge of the eastern lateral moraine, exposing part of the barren zone.

Within the area occupied in 1899 by the long, narrow, southeastern projection of the glacier a stream from the mountain side had built up an alluvial deposit against the ice edge. This deposit now stands exposed as a terrace, cut into by the stream and with the ice contact slope descending so steeply that some of the gravel has slid down since the supporting ice wall was removed by recession of the glacier.

On the steeper slopes of both lateral moraines, near the glacier, stones from the outer

sides were continually rolling down at the time of our visit, because the ice contact slopes of the lateral moraines were too steep for equilibrium after the supporting glacier margin had been removed by melting.

Serpentine Glacier—The Glacier. The Serpentine Glacier (Fig. 45), which descends from the snowfields and cirques (Pl. CXXXIII) on the slopes of Mounts Gilbert and

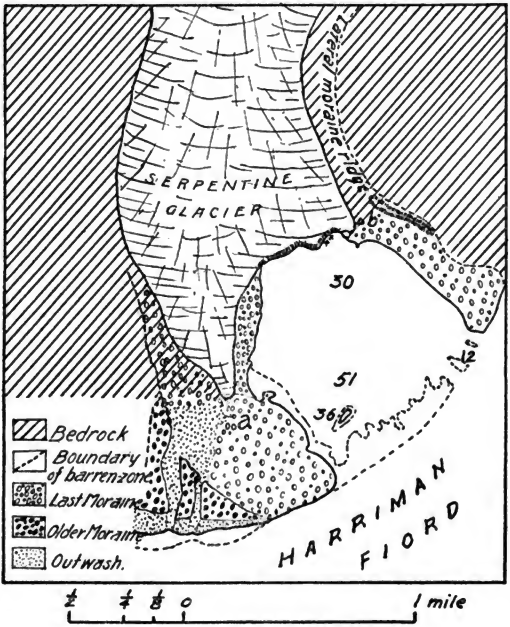


FIG. 45. MAP OF LOWER END OF SERPENTINE GLACIER IN 1910.

Showing moraines, outwash gravels and barren zone. Soundings in feet.

Muir, is about 7½ miles long and ¾ of a mile wide in its mountain valley. At the terminus the width is nearly a mile; and the western half, which ends on the land, projects a half mile farther south than the eastern half, which terminates in a cove of Harriman Fiord. The tidal portion of the glacier terminates in an ice cliff (Pl. CXXXIII) about 2000 feet in length and 200 feet in height, rising out of shallow water a mile from Harriman Fiord. A few small icebergs were discharged, mostly by cascading down the terminus. A rock ledge is exposed beneath the eastern margin.

In contrast to this tidal cliff is the projecting south half of the ice front, a low, sloping surface, with crevasses nearly healed, easy to traverse and thinly veneered with ablation moraine. In front of it is a small area of slumping débris from which much water emerges, proving that part of the flat is underlain by buried ice blocks. Indeed black ice is revealed in places. This projection of the glacier on the western side is due partly to the protection furnished by the morainic cover and partly to the fact that the salt water has undercut the eastern cliff and caused it to retreat faster.

Serpentine Glacier is fed by four or five moderate-sized tributaries from high cirques on Mt. Gilbert, by one from the 6683 foot peak west of Cascade Glacier, and by a large tributary which descends steeply from the slopes of Mt. Muir and enters on the western side about half a mile from the terminus. Farther south on the slopes of Mt. Muir there are also twin cascading glaciers, formerly tributaries of the Serpentine, which now terminate about 3000 feet above sea level.

The white surface of Serpentine Glacier is diversified by a number of medial moraine stripes; and there is a pronounced lateral moraine ridge (Pl. CXXXIII) on the eastern side, left standing by recent retreat. The glacier slopes moderately and is not as severely crevassed as Barry Glacier, although in the last few hundred feet of descent to the water's edge the slope is so steep that travel upon this part of the ice surface was impossible.

Besides the rock ledges beneath the eastern edge of the tidal ice cliff, and the bare rock slopes along the eastern margin of the glacier, there are strong rock ledges near the western margin, one of them cut through by a shallow, smoothly-scoured gorge with vertical walls but with no stream at present. Its upper end hangs in the air, and below its terminus are stony stream deposits.

Slight Change. Between the time of its discovery by the Harriman Expedition in 1899 and our own studies in 1910, the terminus of the Serpentine Glacier has not changed appreciably except in thinning and slight shortening described by Grant and Higgins, on the basis of observations and photographs in 1905 and 1909. Gannett mapped the glacier in a general way in 1899 (Pl. CXXIX, A) and Higgins in more detail ten years later. It is shown upon our 1910 map (Fig. 45).

Moraines. There is clear evidence that not many years ago Serpentine Glacier extended at least a mile farther than now, and that it ended in Harriman Fiord with a tidal ice front 1½ miles or more in length. This is proved by terminal and lateral moraine deposits and by a barren zone near the terminus of the glacier, containing only scattered small shrubs and mosses.

The terminal moraine varies considerably in character, and most of it is free from slumping. The western part is a broad, gently-undulating surface rising perhaps 50 feet above sea level, but being higher and more hummocky on the southeastern border. There is much bare rock, mostly angular and without striae. The eastern portion is a submerged bar extending completely across the mouth of the cove, except for two narrow gaps. At high tide this moraine bar is completely covered; but at low tide nearly the whole of it is exposed. Inside the bar is the cove in which Serpentine Glacier terminates, and in this cove the water is from 30 to 51 feet deep, while at the entrance the greatest depth is only 12 feet. Since the formerly expanded glacier spread out into the bulb shape, it will be seen that the site of this cove inside the crescentic terminal moraine, where there was melting of clean ice without enough débris to fill the shallow inden-

tation may correspond to the interior flats described in several of the Yakutat Bay glaciers.

The lateral moraine on the eastern side of the glacier, mentioned by Gilbert, is one of the best developed deposits of this sort which we have seen in Alaska (Pl. CXXXIII). It is a narrow-crested ridge 30 to 50 feet high and with so sharp a crest that it is difficult to walk upon it. On the western side of this ridge is a bare slope of ground moraine and bed rock extending to the shores of the cove, and to the present border of the glacier; on the eastern side it is bordered by dense mature forest. This lateral moraine is a conspicuous object in the region, curving northward along the margin of Serpentine Glacier for a distance of over a mile, beyond which it could not be seen from sea level. There is a less conspicuous lateral moraine above the western border of the glacier near the terminus, and extending a short distance up the valley. There is also a well developed moraine between the main glacier and the large tributary on the west.

The barren zone, previously noted by Grant and Higgins, comprises all of the area between these lateral moraines, the terminal moraine, and the glacier. It has been covered so recently by the ice that there are only scattered shrubs on the surface, most of the alders being 23 to 27 years old, while the oldest spruce seen had 26 annual rings.

There is an older terminal moraine outside the one described above, the only remnant of it being preserved southwest of the glacier where it is very hummocky, and includes many large bowlders. This moraine rises to a height of approximately 50 or 75 feet and has a wedge-shape (Fig. 45) with the point of the wedge on the shore of Harriman Fiord, whence it broadens westward, in a half mile attaining a width of 2000 feet. Probably this wedge-shape is due to the fact that the eastern portion of the older moraine was destroyed by the recent advance already described. That this outer moraine was made long before the inner moraine is proved by the presence of older trees upon it, there being a thick growth of willow and alder and many spruces and hemlocks nearly a century old. The oldest tree examined was a foot in diameter and had 93 annual rings. The forest on this older moraine is, however, younger and less open than the normal mature forest, such as that outside the lateral moraine to the east of the glacier.

Outwash. The southern side of the older terminal moraine is flanked by a narrow strip, or apron, of outwash gravel and beach deposits, less than one-eighth of a mile wide at low tide, and only 200 to 350 feet wide at high tide. The vegetation proves that most of this outwash gravel flat is associated in origin with the older advance. Most of it is covered with grass, but there are scattered groups of mature trees upon its surface.

Crossing this older terminal moraine, and covering parts of the low, undulating surface of the inner moraine, is a small area of outwash gravel, built from the present ice front. Since until recently streams have flowed over it and it has been undetermined by slumping, there is little vegetation on its surface, the oldest shrubs observed being but 8 years old. Two conspicuous stream-cut channels extend from this inner area of outwash gravels across the older terminal moraine and outwash apron to the sea. They are 1300 and 2000 feet in length, respectively, and where they cross the outer moraine are from 50 to 300 feet in width. Each of them is now abandoned and they were evidently formed at the time of the last advance, when the ice partly destroyed the older terminal moraine and when the glacial streams flowed across the older moraine. In the bottoms of these channels are stream-borne gravels, and abandoned terraces. The scat-

tered young shrubs that grow there contrast with the older trees of the moraine on either side and with the thick grass of the older, outer, outwash apron.

Baker Glacier—General Description. Baker Glacier is a small, steeply-cascading ice tongue, descending from a level of 5000 to 6000 feet in a series of cirques on the southwestern side of Mt. Muir. It has a total length of less than 2 miles (Fig. 46) and terminates about 1300 to 1500 feet above the level of the fiord. At the terminus the width is 3000 feet. The glacier ends on a very steep slope and a narrow pendant ice tongue with a length of 800 feet and a width of 300 feet extends forward from the terminus, its lower end being about 700 feet above the fiord. All of the glacier is white, considerably crevassed, and without moraines.

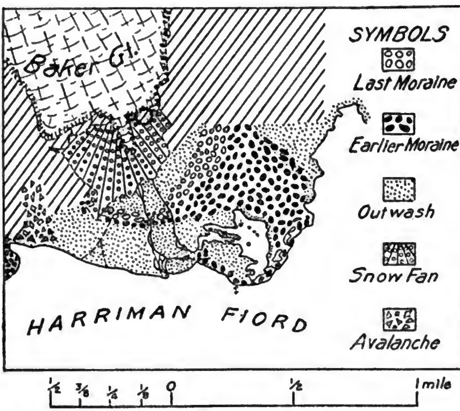


FIG. 46. BAKER GLACIER IN 1910, SHOWING SNOW FAN, MORAINES AND OUTWASH DEPOSITS.

Advance. Photographs by the Harriman Expedition in 1899 and by Grant, Paige, and Higgins in 1905 and 1909 show that, except in one respect, the main features of this glacier have remained unchanged. There was no appreciable change from 1899 to 1905, but a slight advance began between 1905 and 1909, as noted by Grant and Higgins. This advance continued up to 1910, particularly on the north side. Although the form of the glacier front changed very little, the advance was appreciable in photographs.

The Snow Fan. The most conspicuous result of this advance was the increase in discharge of ice from the front of the glacier, which ended on such a steep slope that at the time of our visit blocks of ice were constantly rolling down. In 1905 there was so small a discharge of ice blocks that no accumulation at the cliff base is visible in the 1905 photograph; in 1909 a great snow fan or talus remained until the last of June; in 1910 this was even larger and formed a conspicuous deposit made up of snow and great blocks of ice, some 10 to 15 feet in diameter. When we saw it in July and August, 1910, the accumulation was evidently made up mainly of material that had gathered and remained unmelted during the previous summers, for a large amount of the material between the ice blocks was old snow; but ice blocks were still rolling down and adding to the deposit. This increase in ice block avalanches is clear evidence of an advance, and apparently represents an early stage in the development of a reconstructed glacier; but the deposit

was distinctly not of glacial character when we saw it; it had no motion and was more like a talus deposit than anything else.

Moraine and Outwash. A noteworthy feature, giving to Baker Glacier an interest out of proportion to its present size and activity, is the fact that in former times it has been of greater size, descending the mountain slope and spreading out at its base in a bulb, almost exactly as the Vassar Glacier in College Fiord does at present. Evidence of this expansion is found in the narrow, piedmont coastal plain in front (Fig. 46). This has a total width of 2000 or 3000 feet and is made up of a compound, crescentic moraine, outside which are isolated strips of outwash gravels, while on the inner side is a smaller area of outwash between the terminal moraine and the foot of the snow fan. The two outer areas of outwash are alluvial fans 1000 and 1400 feet wide lying to the right and left of the terminal moraine, which extends to the fiord between them.

The terminal moraine is spoken of as compound because of the evidence of two stages in its building; but the deposit is essentially one continuous crescentic mass of knobs and ridges, hollows with lakes, and undrained areas. The largest of the lakes is over 1500 feet long, but all of the others are very small. The local relief is 20 or 25 feet. This moraine is very stony and in places has no trees, but there are many lichens. The larger part of the outer morainic accumulation is densely covered with forest. Near the inner moraine are willows up to 29 years of age and alders as much as 30 years old. Farther out are mature thickly-set spruces and hemlocks, including trees with 98 to 110 annual rings. The eastern side of this outer moraine extends down to Harriman Fiord, but the western side is separated from the fiord by an alluvial fan of older outwash which we associate with the older moraine because of the age of the trees and shrubs upon it. The older outwash fan on the eastern side supports a dense growth of grass, proving that no glacial streams now flow across it; but there is little other vegetation.

The inner part of the terminal moraine extends in a much narrower crescent than the outer. That it was formed in a relatively recent stage of activity of Baker Glacier is proved by the fact that its irregular surface is not thickly moss covered, as the outer moraine is, and none of the scattered willows and alders are more than 17 or 18 years old.

Outwash deposits are still being made, on the lower slopes of the inner moraine, being supplied with water from the glacier and the snow fan. Some of the streams flow directly down the mountains from the eastern edge of the glacier; but several, after emerging from the glacier, flow down over the rock wall for about 200 feet, then disappear beneath the upper edge of the snow fan, emerging at its lower edge. Two large streams flow over the inner area of outwash gravels and across the terminal moraine hillocks in deeply cut gulleys. Each of these streams is also continuing the building of the outer alluvial fans; but in contrast to the older outwash described above, the gravels now being deposited have no vegetation upon their surface.

Surprise Glacier—Description. The Surprise Glacier (Pl. CXXXIV) has a known length of over 4 miles, and a width at the terminus of 3850 feet. Six tributaries are known (Pl. CXXXV), but since the grade is still low at the most westerly point visible from the fiord, it seems probable that the glacier is even longer and fed by a still larger number of tributaries. The terminal ice cliff is 272 feet high in the middle, and 214 feet high on the northern margin, while back of it is a terminal cascade with a slope of 1900 to 2800 feet to the mile, above which the grade flattens. The glacier surface,

which is severely crevassed from side to side, is stained by one rather strong medial moraine, while a narrow lateral moraine extends along the northern side, and a broad lateral moraine along the southern side.

Retreat. Between 1899, when Gannett mapped Surprise Glacier, and our visit in 1910, the ice front retreated about 6500 feet (Fig. 47). In 1899 the dark-colored southern edge

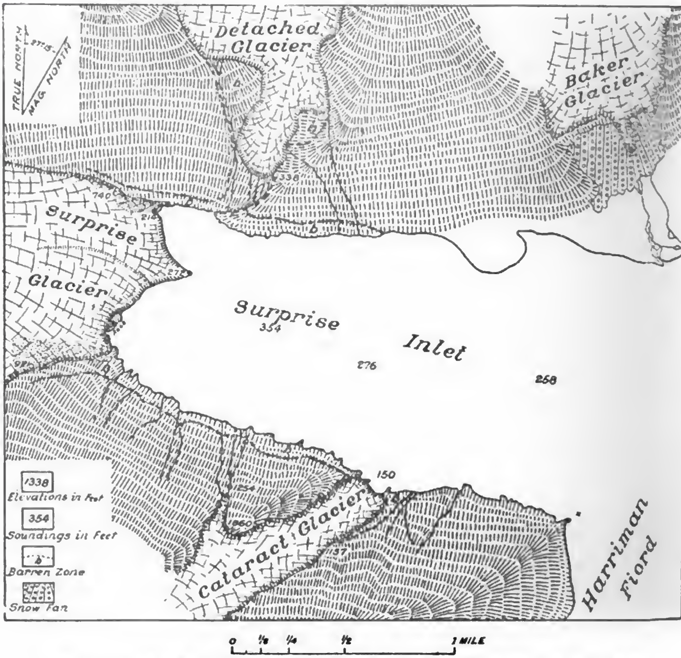


FIG. 47. LOWER ENDS OF SURPRISE, CATARACT, DETACHED AND BAKER GLACIERS IN 1910.

of the terminus just coalesced with the western edge of Cataract Glacier[1] (Pl. CXXIX, A). In 1910 a pronounced cape projected forward a thousand feet, just north of the middle of the glacier front, and on its southern side two rock ledges were visible

[1] This fact, shown both on Gannett's map and the Harriman Expedition photograph, must have been overlooked by Grant and Higgins in their statement that in 1899 the two glaciers were separated a quarter of a mile.

beneath the glacier, the larger one being about 600 feet long. These ledges, seen by Grant and Higgins in 1909 but not in 1905, increased in area from 1909 to 1910, when the glacier seemed to be descending steeply over them. They may mark the head of the fiord.

From this point eastward to Cataract Glacier in 1899 the ice surface was rather flat, and Gilbert noted that it terminated in a lofty cliff.

Barren Zone. A bare zone, whose area increased from 1899 to 1909, as noted by Grant and Higgins, extends eastward on each fiord wall as far as Cataract Glacier and marks the 1899 position of Surprise Glacier. Its upper boundary ascends rapidly westward and in 1910 attained a width of from 300 to 500 feet. Within this barren zone there was almost no vegetation; but there are extensive bare rock slopes on which vegetation will start slowly. The presence of a barren area 150 to 300 feet wide above the present surface of the Surprise Glacier on both sides shows that during the retreat of about 1¼ miles the glacier has also diminished notably in width and thickness.

Detached Glacier. Detached and Cataract Glaciers are cascading tongues formerly tributaries of Surprise Glacier from the north and south respectively. Detached Glacier descends from a cirque on Mt. Muir. It is about 1½ miles long, 3500 feet wide, and has a projecting terminus about 1200 feet wide, which descends on the fiord wall to within a quarter of a mile of the sea, terminating 1338 feet above sea level (Fig. 47). Several streams from various points along the terminus of Detached Glacier, (Pl. CXXXIV) descend the fiord wall in deep-cut gorges. In 1910 there was a moderate sized barren zone around the terminus, across which ice fragments from the glacier occasionally rolled, falling into the alder thickets and into one of the deep gullies below.

The glacier changed very little, if any, between 1899 and 1910, and Gannett's map and several 1899 photographs show that it was already detached from the Surprise Glacier, which at that time extended down the fiord beyond the terminus of Detached Glacier.

Cataract Glacier—Cascading Terminus. Cataract Glacier, which is a mile and a half long and 1100 to 1600 feet wide (Fig. 47), differs from Detached Glacier in still descending the fiord wall to sea level. This is because it occupies a northern slope where melting is minimized and because it has considerably larger snowfields than Detached Glacier. Only its cascading terminus is plainly visible from the fiord and above its hanging valley lip we have not studied it. At the 1860 foot level a northern lobe about 1200 feet in length and 200 to 300 feet in width projects from Cataract Glacier, terminating 1254 feet above the fiord. Several slender streams, with foaming cascades, descend the fiord wall from this terminus.

Cataract Glacier is the only cascading glacier in Harriman Fiord which is now active enough to descend the fiord wall to the sea, as so many slightly-larger cascading glaciers do in College Fiord. The glacier emerges from its hanging valley at about the 2000 foot level, descends the fiord wall (Pl. CXXXV) at the rate of 2850 feet per mile and terminates in a vertical cliff 50 feet or more in height, which rises out of such shallow water that there is practically no undercutting by the sea. Therefore the icebergs from Cataract Glacier are both small and few in number. Most of the glacier surface is severely crevassed, but the eastern edge has a low, sloping terminus, more or less mantled with ablation moraine, over which it is easy to walk. The end of the glacier divides over a low, striated, rock spur.

Advance. Cataract Glacier was beginning to advance in 1910, after having remained

in about the same position from 1899 to 1909, as indicated by comparison of the photographs by the Harriman Expedition and by Grant and Higgins. At the western margin the advancing ice was overriding shrubs of willow and alder, and there were ice blocks sliding from the lateral cliffs into the alders; but higher up on the mountain slope the western margin of the glacier had not yet overridden the whole of its narrow barren zone; and there was still a barren zone on the eastern side extending from 100 to 700 feet beyond the glacier. At the outer edge of this barren zone is a small lateral moraine.

Harriman Glacier—General Description. Harriman Glacier, which is at least 9 miles long and about a mile wide at the terminus, flows northeastward into the extreme southern end of Harriman Fiord. Its whole surface is moderately crevassed, and it is quite free from moraines, there being two medial moraines near the southern side but no lateral moraines. The terminus is a vertical cliff 200 to 300 feet high, from which many icebergs are discharged. Six or seven hundred feet from the middle of the ice front the water is only 90 feet deep.

Névé-Sheathed Slopes. Harriman Glacier differs from most of the others in the Port Wells region in the nature of its tributaries. Instead of being fed by a number of well-defined tributaries, occupying subordinate valleys and each with a length of several miles, it is fed by a considerable number of rather short cascading tributaries, from cirques in peaks 5815 to 8206 feet high. The longest tributary seen from Harriman Fiord is not over a mile or two in length. A great deal of the ice is supplied from névé fields between these short tributaries. These névé fields (Pl. CXXXVI) are intermediate between the snowfields and the ice tongue, and around the borders of Harriman Glacier, large areas of mountain slope are completely sheathed with such névé ice. On the sketch maps by Gannett and by Grant and Higgins these areas of névé are not specifically distinguished from true ice tongues; and it would indeed be impossible to map them accurately without the most detailed work in the névé area.

The phenomenon of ice-sheathed or névé-sheathed mountain slopes, developed on a large scale in Harriman Fiord, is a common feature of Alaskan glaciation of the present time found in other areas, as in Passage Canal, along the eastern side of Lynn Canal in southeastern Alaska, and doubtless in many other localities where the snowfall is great and the snow line is rather low. If the snowfall should decrease slightly it seems likely that these névé-sheathed slopes would become bare and that the short tributaries would shrink and become better-defined, though smaller, ice tongues than at present.

Retreat. The front of Harriman Glacier has oscillated considerably since 1899 when it was described by Gilbert and mapped by Gannett. Previous to that year there seems to have been some retreat, for Gilbert observed that the glacier was not "closely approached by forest growth, but shrubs were seen on the shore of the fiord within a few hundred yards of the ice." This suggests a narrow barren zone.

Grant and Higgins state that "photographs of the eastern side of the front of Harriman Glacier taken from Point H (Fig. 50) in 1905 and 1909 show that this side of the glacier (eastern) retreated approximately 700 feet between these dates. A comparison of an 1899 photograph with the above indicates that between 1899 and 1905 the east side of the glacier retreated about half the above distance. As the two photographs were not taken from the same point this estimate of the retreat between 1899 and 1905 is only approximate. On the west side of the ice front a careful examination in 1909 of the glacier from the position of a photograph taken in 1899 showed no noticeable difference

in the position of the glacial front. In 1899 a considerable embayment existed in the eastern third of this glacier, but was not present in 1905 and 1909."

Advance. It is clear that there was more or less continuous retreat of Harriman Glacier during the decade following its discovery; but between 1909 and 1910, an advance was instituted. This was clearly shown on August 1, 1910, when a comparison with the glacier front of the Harriman Expedition photograph, alluded to by Grant, showed an advance of the western margin of 700 feet (Pl. CXXXVII), accompanied by great thickening at the terminus. The western gravel bar, shown in another Harriman Expedition photograph, upon which the margin of the glacier terminated in 1899, was not visible at the time of our only visit, which was near high tide while the 1899 picture was taken when the tide was part way out. On the land near by, however, we saw the glacier advancing and destroying alder by overriding and by sliding down of ice blocks.

The 1909 observations of Grant and Higgins make it certain that the change from retreat to advance came between 1909 and 1910 and that the whole of the 700 feet advance was during the last year. The eastern margin also advanced, coming forward the whole distance that it had retreated from 1899 to 1909. In 1910 it seemed to be slightly beyond the 1899 position and there was much thickening of the eastern terminus, which Gilbert says was "low and irregular" and rested on a "detrital bank" in 1899, which was not visible in 1910. There seems to have been far more iceberg discharge in 1910 than in 1899.

In addition to the advance and thickening of the part of the eastern margin on the shores of the bay, the glacier was advancing on the land. Annual plants in the barren zone were being buried and a ridge of push moraine a foot or two high lay at the base of a lofty, uncrevassed ice cliff. On August 1, this ice cliff was 158 feet from an older terminal moraine which marked the maximum of a former advance, doubtless before 1899.

The oscillations of Harriman Glacier during the period of observation may be summarized as follows.

Date	*Oscillation*	*Amount*	*Observer*
Before June 27, 1899	Probably retreat	———	Gilbert
June 27, 1899, to Aug. 20, 1905	Retreat	About 350 feet	Grant and Paige
Aug. 20, 1905, to June 29, 1909	Retreat continued	700 feet	Grant and Higgins
June 29, 1909, to Aug. 1, 1910	Advance	700–1050 feet	Martin

Dirty Glacier. There are a number of smaller ice tongues in southern Harriman Fiord, including one or two in hanging valleys above Harriman Glacier (Pl. CXXXVII), of which they are disconnected tributaries, and eight or more farther north in the fiord which were formerly tributaries of the extended Harriman Glacier.

Of these ice tongues the Dirty Glacier, on the eastern side of the fiord, just north of Harriman Glacier terminus, is exceptional in having ablation moraine cover over a large proportion of the terminus. Its névé fields coalesce with those of Harriman Glacier, and although it has a steeper grade than the Harriman it contrasts strongly with neighboring

cascading ice tongues, like Roaring Glacier, for it slopes at an angle of only about 12°. It ends 100 feet or less above the level of the fiord, from which it is separated by an outwash gravel fan, and by a morainic area with many small ponds. There was practically no change in this glacier from 1899 to 1910, but in former times it evidently expanded and formed a bulb glacier, and still earlier was tributary to Harriman Glacier.

Roaring Glacier. Roaring Glacier, three-fourths of a mile north of Harriman Glacier, on the northwestern side of the fiord, is spoken of by Gilbert as an ice tongue which "owes the peculiarity suggesting its name to an abrupt change of grade. From a comparatively gentle slope it passes to one so steep that loose masses find no lodgment, and as its movement steadily projects its end beyond the point of inflection, fragments of ice break away and tumble down the steep incline, to gather in a heap far below, where they lie until melted."

It seems to have been less active in 1905 than in 1899, for a photograph by Grant and Paige shows that in 1905 there was no heap of ice fragments below the glacier, but that, as in the case of Baker Glacier in the same year, there was an extensive area of bare rock below the terminus.

Roaring Glacier became more active between 1905 and 1910 for the barren zone, which was free from snow in August, 1905, was partly covered in August, 1910, by an extensive snow and ice block fan, similar to that of Baker Glacier. This fan could only have been formed by an increase in activity of the glacier with accompanying sliding down of many ice blocks from the terminus, which is 700 to 1000 feet above the fiord. Advance seemed to be still going on in August, 1910.

Roaring Glacier has evidently descended to fiord level since it became independent of Harriman Glacier, for around the borders of the snow fan there is a crescentic terminal moraine, the outer part of which is covered with shrubs. It extends down to the fiord, where it makes a small cape; and since it rises steeply from the water's edge, it is evident that the glacier reached tidewater when last expanded. The crescentic character of this deposit shows that it was not built by Harriman Glacier, but the knobby morainic topography does not absolutely preclude the possibility of its having been built by an avalanche rather than by a bulb-shaped extension of Roaring Glacier, though we do not think this avalanche origin likely.

Wedge Glacier. Wedge Glacier lies on the eastern side of the fiord, opposite Roaring Glacier (Fig. 50). Its general shape is indicated by its name, forming a striking contrast with ice tongues, like Dirty Glacier, that have bulb or fan-like termini. Wedge Glacier ends in a narrow valley a few hundred feet above the fiord, and except in detached spots was covered with snow clear to the end in August, 1910.

Smaller Glaciers. Two unnamed ice tongues north of Wedge Glacier have snowfields which coalesce with that of Toboggan Glacier and terminate ¼ and ¾ of a mile respectively from the fiord. Three small ice tongues on the western side of the fiord between Roaring and Cataract Glaciers, together with Roaring Glacier, two glaciers in hanging valleys above the northwestern end of Harriman Glacier, and several others, fall in a class described by Gilbert. He says, they "occupy elevated valleys far above the main trough, and these upland valleys probably constitute a system initiated at an earlier epoch, when the fiord was flooded with ice to a greater depth. . . . No measurements were made, but it is evident from an inspection of photographs that the heights of such features in this neighborhood are approximately the same as in the vicinity of

Serpentine and Surprise Glaciers, and it is possible that a number of minor glaciers observed on both sides of the fiord constitute with these a general system."

Toboggan Glacier—The Ice Tongue. Toboggan Glacier, on the eastern side of Harriman Fiord, nearly opposite Serpentine Glacier, descends the fiord wall from a snowfield of unknown extent, from which, three miles to the south, two smaller ice tongues also project. The valley portion of this cascading glacier is about a mile long and from an eighth to a quarter of a mile wide. Below the lip of its hanging valley it has not as steep an ice cascade as the Cataract Glacier and the cascading glaciers of College Fiord have, although it has a slope averaging 12° to 14°, steepening to 29° to 31° at the end. The glacier is clean, except for a medial moraine of black slate, near the northern border, and two small V-shaped areas of ablation moraine at the terminus.

At the terminus (Fig. 48) is a ledge of black slate 12 to 15 feet high, lying exactly at the northern end of the medial moraine, and the main glacial stream emerges from an ice cave immediately west of this. The glacier partly covered this ledge in 1910, sloping back evenly from its top. West of this ledge is a low slate outcrop beneath the terminus of the glacier. One isolated morainic hummock, 15 or 20 feet from the glacier, and just west of the stream, still had ice in it and was rapidly slumping. The rest of the ice front seemed to rest upon the edge of the gravels, nowhere underlying them, and causing slumping through melting.

The Outwash Gravels. In 1910, Toboggan Glacier terminated just inside its mountain valley, a little over a quarter of a mile from the fiord (1600 feet), at an elevation of about 112 feet above sea level. Between it and the fiord was a very symmetrical outwash gravel plain, projecting into the fiord for about an eighth of a mile as a delta. Above the surface of this outwash fan, which slopes at the rate of 370 feet to the mile, rise a number of ledges of bare rock, and also several small, barren, gravel hillocks and one tree-covered gravel knob. Some of the gravel mounds are remnants of an older dissected outwash plain, while others may be remnants of older terminal moraine built when, as Grant and Higgins [1] believe, the glacier reached practically to tidewater. One large stream flows over the fan near the middle, and one small one on the extreme southern edge. The main stream has shifted slightly since 1905. There are also older outwash gravels, described in a later section.

Advance and Retreat. Grant, Paige, and Higgins visited Toboggan Glacier in 1905 and 1909, determining an advance of 400 feet and then a retreat of about 650 feet between these dates. They found that "in 1905 the center, or most advanced part of the glacier, was 723 feet, as determined by pacing, distant from the cairn (at a, Fig. 48). . . . Just at the extreme front of the ice at this date was a low rock ridge crossing the valley. In 1909 the most advanced part of the glacier was 252 feet farther back than in 1905. However, in 1909 a freshly deposited low moraine in the northern half of the plain in front of the glacier indicates that the ice sometime between 1905 and 1909 had been about 400 feet in advance of its position at the earlier date."

A photograph by Grant and Paige (Pl. CXXXII, B) shows that the glacier was much more crevassed in 1905 than in 1910, and that during its retreat the terminus has narrowed slightly on the southern side. It was still receding in July, 1910, having retreated 75 feet at the point of Grant's measurements. Future measurements of the

[1] Bull. Amer. Geog. Soc., Vol. XLIII, 1911, legend of Fig. 13, p. 337.

retreat of Toboggan Glacier will be facilitated by the black slate ledge at the northern end of the medial moraine. A good point from which to measure, in case of advance of the glacier, is the nearest graywacke ledge to the north (e, Fig. 48) which was 517 feet from the ice front in 1910, and is a low roche moutonnée, beautifully striated. There are other graywacke ledges further west.

The low moraine described by Grant seems to have been destroyed by the glacial streams between 1909 and 1910.

In front of the southwestern portion of the Toboggan Glacier, an area of parallel furrows and minute ridges similar to those seen by Gilbert at Columbia Glacier, had been exposed by melting since 1905. They extended north and south, disappearing under the edge of the ice, and were made of till and outwash gravel containing fragments of finely-shredded wood. About 70 feet from the ice front, and parallel to it, was

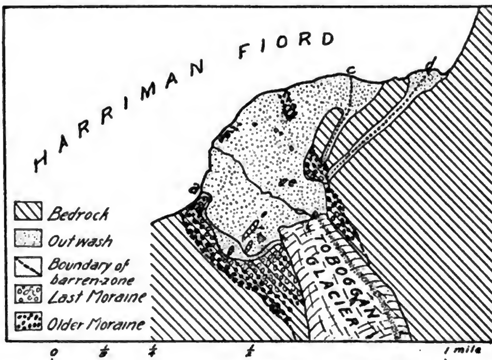


Fig. 48. Map of Lower End of Toboggan Glacier in 1910.
Showing outwash delta, moraines and abandoned stream gorges.

a diminutive ridge, 6–12 inches high, crossing the erosion furrows nearly at right angles. It suggests either a crevasse deposit or a minute terminal moraine, probably the former.

Abandoned Marginal Gorges. In former times Toboggan Glacier was undoubtedly a tributary of the large ice tongue of Harriman Fiord, and at a later stage, though disconnected, it was both longer and wider than now, as is shown by the barren zone about its borders. On the eastern side of the glacier, outside this barren zone, are two abandoned stream courses (c and d, Fig. 48), occupied by marginal drainage at a time when the glacier was more expanded than now, but later than the period when it extended up to the barren zone.

Barren Zone. A barren zone around both the eastern and western borders of Toboggan Glacier bears sparse vegetation which furnishes evidence of two stages in the history of the glacier. Outside the barren zone is mature, close-set forest, with thick moss but without undergrowth, the trees, which are from 11 to 19 inches in diameter, probably

being at least a century or two old. They end along a straight line but there is no terminal moraine or line of dead tree trunks at the forest edge.

The outer part of the barren zone, farthest from the glacier, is about 100 feet wide and bears scattered trees 65 to 70 years old; while the inner zone, nearest the glacier, has only a very few scattered shrubs none of them more than 20 years old. There is not only a difference in age of trees but a difference in the amount of moss on the rocks in the younger and older parts of the barren zone. Parts of these zones have morainic soil, parts are bare rock and in places there are small stony lateral moraines containing dead wood and rolls of peaty soil.

Growing upon the surface of the largest gravel knob in the outwash fan (b, Fig. 48) is a dense thicket of alders up to 30 years old and some hemlocks, of which the oldest, observed in 1910, was 72 years old. The deposit in this knob is, therefore, correlated with the advance which extended up to the edge of the barren zone, as are also (a) a high level delta in the small valley on the western side of the alluvial fan, and (b) some small mounds of gravel which may represent older outwash. The high level delta which stands about 50 feet above sea level, has a gently-sloping surface and seems to have been built into a small marginal lake at the outer edge of the barren zone. On its surface are trees about 70 years old and it is trenched deeply by the present stream course.

During this period of advance the glacier seems to have had a pronounced bulb shape and to have extended to the fiord; but the second advance evidently did not go nearly so far. Most of the main delta surface is barren, being still occupied by shifting streams, in contrast with (a) the delta at the end of the eastern gorge (d, Fig. 48) on which grow 70 year old conifers, and (b) the delta at the end of the later (western) gorge (c, Fig. 48) on which we saw no alders over 20 years old. In the zone where this later delta coalesces with the present delta of Toboggan Glacier streams there were scattered trees estimated to be a century old, some of them standing dead, some living but in process of being buried and killed by recent gravel deposition. Two or three conifers also grow at sea level on the extreme western edge of the main delta; but none of these appear to be over 60 or 70 years old.

Port Wells

General Description. Port Wells (Fig. 49), extending north-northeast from Passage Canal to Pt. Pakenham at the entrance to College Fiord and Barry Arm, is 15 miles long and 6 miles wide. On the western side are Pigot and Bettels Bays, each about three miles long, and four shorter, subordinate bays. The eastern wall is interrupted by two small bays and also by the entrance to Esther Passage, which separates Esther Island from the mainland. A low rock rises above sea level at all stages of tide near the southeastern entrance to Port Wells a mile and a quarter off shore; but with the exception of bars at the entrances to College Fiord and Barry Arm the rest of the fiord is deep. The fiord walls of the mainland rise steeply, but nowhere so precipitously as in upper College Fiord and in Harriman Fiord.

Small Glaciers of Port Wells. The present glaciers of Port Wells are the Bettels and Pigot Glaciers, two moderate-sized ice tongues each terminating a little over a mile from the head of the bay of the same name. There are also a number of much smaller glaciers in the mountains to the west of the entrance to Barry Arm.

A group of three small glaciers on the western side of Barry Arm, near its junction with Port Wells, occupy the upper part of a large cirque, the hanging valley lip of which is approximately 100 feet above sea level. There are still smaller ice remnants on the mountains between Port Wells and Harriman Fiord. The general extent of all these small glaciers was sketched by Grant and Higgins in 1909.

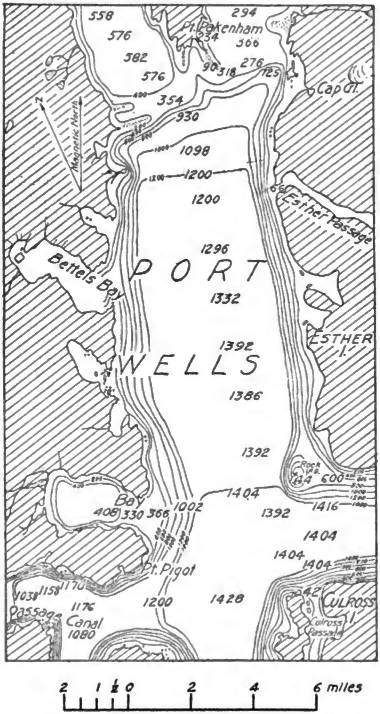


FIG. 49. SUBMARINE TOPOGRAPHY IN PORT WELLS.

Soundings in feet. Submerged contour interval 200 feet (Outline of coast after Grant and Higgins).

Bettels Glacier, which sends a stream to Bettels Bay (Fig. 50) heads in the same snowfields which feed Harriman and Dirty Glaciers. It is about 4 miles long, ½ mile wide, and terminates about 1½ miles southwest of the head of Bettels Bay.

Pigot Glacier (Fig. 54) heads in cirques and snowfields adjoining those of Bettels Glacier and of several small glaciers in Passage Canal. It is about 3 miles long, ½ mile wide, and ends about 1½ miles from the head of Pigot Bay. It had about the same position upon Applegate's map in 1887, Glenn's map in 1898, and Grant and Higgins' map in 1909. The glacier surface is clean, apparently little crevassed, and is striped with four prominent medial moraines. A good-sized tributary enters from the east, another from the north, and another from the south not far from the terminus, the latter cascading down to the main ice tongue, with so steep a grade that it has the appearance of being an avalanche where the ice is heaped upon the main Pigot Glacier, although it probably has no avalanche relationship. Near it are four or five small ice tongues, ending high on the mountain slopes which were formerly tributaries of Pigot Glacier. None of these glaciers have ever been studied at close range.

Glacial Erosion

Glacial Erosion above Sea Level—General. The evidence of intense glacial erosion in Harriman Fiord, Barry Arm, and Port Wells, is plainly visible above sea level and clearly indicated by soundings. Throughout the fiords, glacial striae, conforming in direction to the different trends of the fiords, and roches moutonnées are present upon bare rock ledges near sea level, being especially well shown in the barren zones. These features prove clearly that ancestors of the present glaciers completely occupied the fiords, and suggest the probability that the fiords were given their present straight courses, simple outlines, and steepened lower slopes by ice erosion without sinking of the land.

The extreme height of glaciation in Harriman Fiord and Port Wells was estimated by Gilbert to be 4000 feet or more, on the basis of the "cordon of high hanging valleys" surrounding Harriman Fiord. "Above Barry, Serpentine and Surprise glaciers they contain hanging glaciers at a general height of about 4000 feet, and east of Harriman Glacier their ice banks coalesce in a continuous terrace along the valley wall. The surface of the trunk glacier to which they are adjusted probably lay 5000 feet above present sea level."

The evidence from sea level indicates that the 2900 foot hill back of Pt. Doran was completely overridden, for the summit and slopes are rounded, and the cliffs are all of such small size as might readily be made by glacial plucking or by post-glacial weathering. The same is true of the summit of Esther Island east of Port Wells. There is a general rounded appearance to all the bare summits throughout Harriman Fiord and Port Wells below 5000 feet, except near the present glaciers where there are many cirques separated by narrow arêtes and sharpened peaks.

Plucking and Oversteepening. Some of the lower slopes have been steepened to precipices. There are many illustrations of steepened slopes in Harriman Fiord and Barry Arm. Oversteepening of the lower slopes of the fiord wall between Baker and Detached Glaciers recently caused a great avalanche.

Because of the steepness of the slopes and the consequent absence of soil, many areas of bare rock on the fiord walls are as yet without vegetation excepting in joints and hollows. This condition is well illustrated on the south shore of the inlet leading to Surprise Glacier where the roches moutonnées, glacial grooves, and striae are still well preserved. Many of the beautiful waterfalls in Harriman Fiord and Barry Arm, as along the coast between Pt. Doran and Toboggan Glacier, cascade down bare rock slopes made smooth by glacial erosion. The bare knob at the extreme southeastern limit of the barren zone on the eastern side of Barry Arm shows the roches moutonnées form very well indeed.

Asymetrical Fiord Walls. Pt. Doran is a gently sloping spur which contrasts with the much oversteepened opposite slope below Mt. Curtis, because the united Barry-Harriman Fiord Glacier flowed faster and eroded more efficiently on the eastern side of this curve. That it did not fail to erode the hill behind Pt. Doran also is proved by the fact that all rock ledges not covered with soil or moss are rounded knobs with striae and grooves, a good example of which is seen in the great, overhanging glacial groove shown in Pl. CXXXVIII.

The eastern and western sides of Port Wells form a striking contrast, the former having a markedly oversteepened lower slope, rising up to a distinct shoulder, while the western

side has long sloping spurs not truncated excepting at the tip, nor oversteepened so much as the eastern side. This might possibly be explained by a difference in the rock,[1] for part of the eastern shore is granite, while the western coast is all slate and graywacke, were it not for the fact that southern Esther Island and the coast north of Esther Passage, on the eastern shore, are made up of slate and graywacke, and yet these parts are nearly as much oversteepened as the granite coast. The probable explanation is that there has been differential glacial erosion on the opposite sides of the fiord, the former Port Wells Glacier probably hugging the eastern shore because many tributaries entered from the opposite side.

Cirques and Hanging Valleys. Another evidence of profound glacial erosion comes from the cirques and hanging valleys, of which there are many. There is a well-developed hanging valley, for example, on the western side of Barry Arm, about a mile south of Cascade Glacier (Pl. CXXXIX). The upper part of this valley is still occupied by a small glacier from which a stream descends to the fiord, first with a moderate grade, then precipitously over the lip of the hanging valley which lies between 1000 and 1100 feet above sea level. The stream cascades down to Barry Arm in a large waterfall and a filmy bridal-veil fall, plunging nearly 500 feet in one vertical cascade and altogether descending 1050 feet with horizontal flow of less than 500 feet. This hanging valley has only recently been revealed and the cascades existed only a few years, for in 1898 the Barry Glacier extended up to the lip of the hanging valley. The hanging condition was partly revealed in 1899 but not fully until some time between 1905 and 1908 when the terminus of Barry Glacier had retreated northeast of the site of the waterfall so that the full height of the hanging valley above sea level was exposed.

Besides this typical hanging valley there are many others, of which those containing Cascade, Cataract and Toboggan Glaciers, and the small ice tongues southeast of Barry Glacier may be cited as examples. There are also many cirques from which the ice is completely melted, as between Cascade and Serpentine Glaciers, and on the slopes of Mt. Muir.

Glacial Erosion below Sea Level—Longitudinal Profiles of Fiords. The depth of water at the fronts of the tidal glaciers (Barry 426 feet, Surprise 354 feet, Cataract 150 feet, Harriman 90 feet, Serpentine 51 feet) proves that these ice tongues are all resting on the bottom of the fiord and are therefore still engaged in eroding it.

Within Harriman Fiord the depth of water varies from 90 to 510 feet (Fig. 50), and the longitudinal profiles are undulating, due either to glacial basining or to complication of glacial deposits, a point discussed below. The fiord bottom slope, from Pt. Doran to Pt. Pakenham, increasing evenly in depth from 414 to 576 feet, averages 32 feet to the mile, and this slope is not interrupted by basins due to erosion, or by note worthy submerged glacial deposits.

At Pt. Pakenham where the fiord is shallowed locally by what is interpreted as a submerged moraine, the depth increases suddenly from 576 to 930 feet (Pl. CXLI). As this is at the junction of Barry Arm and College Fiord we infer that the former expanded Port Wells glacier, which was fed from these inlets, was able to erode more efficiently than either the smaller Barry Arm or the College Fiord glaciers, but as they were subequal in size they have produced a confluence step at the north end of Port Wells. The longitudinal profile is interrupted by a step at the mouth of each fiord, however,

[1] See geological map by Grant and Higgins, Pl. II, Bull. 443, U. S. Geol. Survey, 1910.

because of increase in glacial erosion by the united Barry Arm and College Fiord Glaciers. These submerged steps are apparently the same as the rock treads produced by differential glacial erosion on the land. The lip of the submerged tread of the Barry Arm step is 354 feet higher than the bottom of Port Wells; but if the ridge interpreted as a submerged moraine is added the discordance is 576 feet in all.

The next three soundings in Port Wells, a mile apart, show depths of 1098 to 1200 feet (Fig. 49); and the following seven soundings, two miles apart, show depths of 1296 to 1404 feet. The bottom slope of Port Wells, therefore, averages 28 feet to the mile, being a little flatter than the bottom slope of Barry Arm (Pl. CXL) above the step.

Cross-Sections of Fiords. The cross-sections of southern Port Wells and outer Passage Canal show a fiord bottom 1400 feet deep and between two and three miles wide, bordered by sides sloping 1000 to 1200 feet to the mile. Such a cross-section is characteristic of a glacially-eroded fiord, and is repeated in narrower sections throughout Port Wells, Barry Arm, and Harriman Fiord.

Submerged Hanging Valleys. Pigot Bay is a submerged hanging valley (Fig. 49), with water 330 to 408 feet deep, and with the lip 1038 feet above the fiord bottom. Bettels Bay and the small indentations on the western side of Port Wells may be presumed to have a similar discordant relationship though no soundings were made. The indentation south of Bettels Bay is barred at the mouth by rock islands, and if the channels between are shallow, as is probable, the discordance of grade here is over 1300 feet. Submerged hanging valleys are also suspected in the coves on the eastern side of Port Wells south of Esther Passage. This passage itself has a hanging relationship to Port Wells, the water being 66 feet deep at its entrance, while the main Port Wells fiord is 1200 feet deep.

A large valley on the eastern side of Barry Arm joins the fiord at sea level, and if there are rock ledges near the mouth it hangs over 500 feet above the fiord bottom.

The deposits near the mouth of Surprise Glacier inlet make it impossible to determine the erosion relationship of Surprise and Harriman Glaciers; but, though this cannot be stated positively, it seems probable from the existing depths that the inlet of Surprise Glacier hangs 100 feet or more above that of Harriman Glacier. However, the cove in which Serpentine Glacier terminates, is clearly a submerged hanging valley, the depth of water being 12 to 51 feet near the glacier front, and 474 feet in the fiord opposite, so that the total amount of discordance is at least 423 feet (Fig. 50).

Glacier Deposits in Harriman Fiord and Port Wells

Partly as a result of the steepness of the fiord walls, the glacial deposits in these fiords are not extensive. They include a veneer of ground moraine on all but the steeper slopes, small belts of lateral moraine on the fiord walls, some areas of outwash plain, and some terminal moraines above and below sea level.

Moraine. The areas of ground moraine call for no particular description. They consist of till and bowlders of various sorts, including rock carried out from the mountains and not present in the country rock of the region near where they lie. There are doubtless deposits of similar material and of fine silts below sea level in the fiords, but of the nature and distribution of these nothing is known.

The lateral moraines are especially well developed near some of the existing glaciers,

forming pronounced, saw-tooth ridges at the borders of several of the barren zones, as already described.

The only well defined moraines above sea level recognized in this region are the crescentic moraines in front of Baker and Roaring Glaciers and fragments of terminal moraine in front of Serpentine (Fig. 45) and Barry Glaciers (Pl. CXL). A valley on the eastern side of Barry Arm contains Lake Cecelia, which Mendenhall described in 1898[1] as held in by "a low morainic ridge" which "may be the terminal of the glacier now in retreat at the head of the valley, or may be a lateral left by a tributary ice stream which enters from the east." In other places the terminal moraines either have been cut away by streams or were not built because the glaciers terminated in the sea, or were retreating rapidly.

Outwash. The areas of outwash plain have already been described, the one in front of Toboggan Glacier (Fig. 48) being the largest. There are similar areas near the Serpentine and Baker Glaciers, on the western side of the terminal moraine which was being built by Barry Glacier in 1898–99, and in front of Dirty, Pigot, Bettels, and several of the smaller ice tongues.

Submerged Moraines. Of the terminal moraines below sea level the most conspicuous one is a continuation of one on the land. It occupies the position of Barry Glacier in 1898 and 1899, has a roughly crescentic form (Pl. CXL) and rises to sea level in three localities. At all other points it lies within from 96 to 234 feet of the surface, with deeper water on both the inner and outer sides. Opposite Pt. Doran, where this terminal moraine rises to the level of the sea, moderate-sized banks occur at low tide, the one nearest the point being about 1200 feet long, and made of till and bowlders. It is covered at half tide. Besides these three banks there is a shoal in the middle of Harriman Fiord, where the water is only 18 feet deep at low tide; and near the large isolated ice mass, on the western side of Barry Arm, a projecting peninsula 1800 feet long is exposed at low tide, continuing the belt of terminal moraine above sea level. The Barry Arm portion of this crescentic moraine nowhere rises to sea level, the water upon it being 120 to 234 feet deep. Between this submerged terminal moraine and the front of Barry Glacier, the water increases rapidly in depth, being 426 feet deep half a mile from the ice front, and increasing in depth at about the same rate westward up Harriman Fiord and southward down Barry Arm.

The submerged continuation of the terminal moraine of Serpentine Glacier is shown in Fig. 45. It is a rocky bar, practically all exposed at low tide, and rising 39 feet or more above the bottom of the cove in which the glacier now ends.

Another shoal area in Harriman Fiord, which might be interpreted either as terminal moraine below sea level or as rock-swell due to differential glacial erosion, is at the junction of the inlets which lead to Surprise and Harriman Glaciers (Fig. 50). Here there is an irregularity of the fiord bottom rising nearly 300 feet above the bottom of the fiord to the south, the crest of it being 90 feet below sea level in mid-fiord. If morainic, it was built by Harriman Glacier at a time when it extended 6 miles farther north. Icebergs strand upon this shoal more than ¼ mile offshore on the western side of the fiord, suggesting a moraine bar here.

North of this, opposite Baker Glacier, the water is 120 to 174 feet in depth, being

[1] Mendenhall, W. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, p. 325.

therefore shallower than is normal in the fiord. The association of this shoal with a broad point between Baker and Detached Glaciers (Fig. 47), which projects 1150 feet into the fiord and has morainic topography, suggests a crescentic moraine built here by the Serpentine Glacier. The water deepens rapidly to 258 feet west of this shallow area, and to 282 feet east of it. Our soundings are not detailed enough to determine whether or not these 90 and 120 foot irregularities represent a continuous submerged moraine built by the formerly coalescing Harriman and Surprise Glaciers, or indeed to show conclusively that they are moraines rather than irregularities of the fiord bottom due to differential erosion.

At the northern end (Pl. CXLI) of Port Wells there are several sets of shoals which are best interpreted as submerged terminal moraines, or moraine bars, of the sort seen in Unakwik Inlet, but with less of their length exposed at low tide. They span the mouths of Barry Arm and

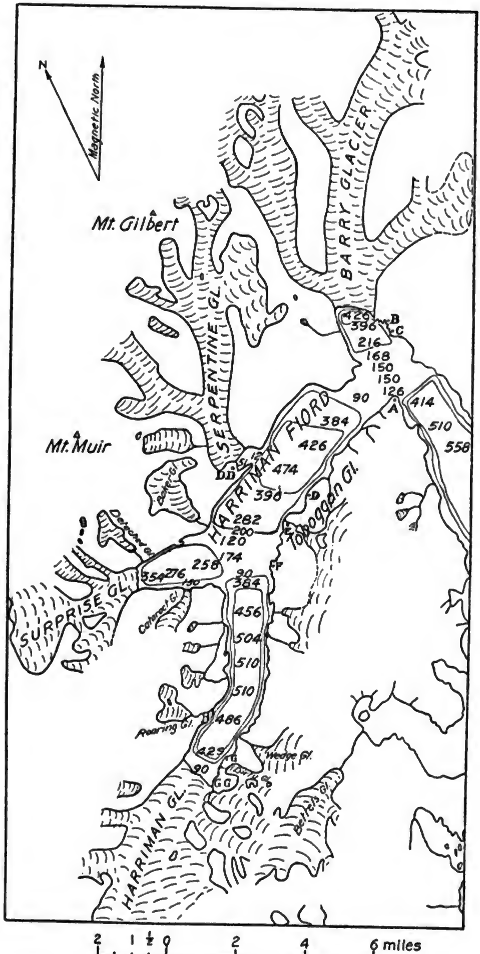


FIG. 50. SUBMARINE TOPOGRAPHY OF HARRIMAN FIORD.

Soundings in feet. Submerged contour interval 200 feet. Photographic Stations A, DD, etc. (Outline of shores and glaciers after Grant and Higgins.)

College Fiord at Point Pakenham, where these fiords unite to make Port Wells. The one at the mouth of Barry Arm has 354 feet of water in mid-fiord but rises 222 feet higher than the bottom a mile farther up the fiord. Its ends are marked by spits running out from each shore $\frac{3}{4}$ and $1\frac{1}{4}$ miles respectively, where alongshore currents could not possibly have built spits without a shallow foundation, such as a moraine to guide them. They are, moreover, composed of morainic material too large to have been carried and built in this specialized form by anything but ice. The one on the western shore is double, suggesting two halts of the expanded glacier.

The southernmost of the two submerged moraines at the entrance to College Fiord has 276 feet of water in mid-fiord, rising 90 feet above the fiord bottom half a mile to the north and 450 feet above that to the south. Its visible portion consists of a moraine bar extending a mile and three-eighths from Pt. Pakenham on the west; but there does not seem to be any similar bar extending out from the eastern shore. The material in the western bar includes coarse angular bowlders, and the part nearest shore supports clumps of large trees.

A mile and a quarter north of this is another submerged bar lying at a depth of less than 294 feet. It rises approximately 50 and 70 feet above the fiord bottom to the north and south and has a point visible at low tide, extending out toward it from either shore.

These three submerged bars, interpreted as moraines, seem to have been built just after the Port Wells Glacier had retreated far enough to separate the Barry Arm and College Fiord ice tongues; and the double character in each case suggests either two halts or a readvance.

A moraine similar to these is possibly being built at the present terminus of Harriman Glacier, for:— (a) there are no rock ledges exposed, as at Surprise Glacier, to suggest that the head of the fiord has been reached; (b) a partial shoal fronted the terminus of the glacier in 1899; and (c) six or seven hundred feet from the ice cliff of 1910, after 700 to 1050 feet of advance, the water was only 90 feet deep, in striking contrast to the depth near the front of other large tidal glaciers in Prince William Sound. Glacial streams as well as ice borne detritus, are doubtless contributing to it.

Relationships of Vegetation to Glacial History

The vegetation in Port Wells, Barry Arm, and Harriman Fiord furnishes much evidence of the recent history of the region. There is thick coniferous forest on the lower slopes of Port Wells and Barry Arm up to Pt. Doran. Above the barren zone of Barry Glacier it extends continuously up to Cascade and Coxe Glaciers; and, even beyond these cascading glaciers, vegetation extends for a short distance, the alder thickets extending farther than the spruce and hemlock. Barren zones around Serpentine, Surprise, and Toboggan Glaciers also interrupt the continuity of the coniferous forest, which extends southward to Baker Glacier on the western side of Harriman Fiord. There is one small clump of conifers between Baker and Detached Glaciers 600 or 700 feet above sea level, west of which alder thickets extend to Surprise Glacier on the northern shore of that inlet. Alders grow between Surprise and Cataract Glaciers on the southern side of Surprise Inlet, and alders and two stunted spruces are found between Cataract Glacier and Harriman Fiord. The eastern side of Harriman Fiord has thick spruce and hemlock

forest from Pt. Doran to Wedge Glacier and alder thickets extend to Harriman Glacier, contrasting with the western side which has alder thickets between Surprise Inlet and the head though only one very stunted spruce was seen.

The presence of thick forest nearly at sea level on Point Doran, with trees possibly over a century of age, shows that Barry Glacier has not overridden this point for at least 100 years, as Gilbert pointed out; and it seems probable that since it ceased to coalesce with Harriman Glacier, Barry Glacier has never reached Point Doran and dammed Harriman Fiord into a lake; for there are neither elevated shorelines nor deltas along its margins. Excepting in southern Harriman Fiord the forest is mature and vigorous, and there are fewer open glades in the coniferous forest of Harriman Fiord than near Columbia Glacier. The oldest tree whose age we determined was 22½ inches in diameter and had 225 annual rings. Muir, however, found one hemlock in 1899 [1] only nine inches in diameter which had attained 325 years. We found trees in Harriman Fiord in 1910 which judging by their diameters, must have been even older than that noted by Muir. Fernow observed in 1899 [2] that logs in Prince William Sound showed the following relationships of diameter to age:

"50 annual rings, 11 inches			
72	"	"	12 to 15 inches
80	"	"	20 inches
125	"	"	22 inches."

This vegetation proves clearly that these fiords have not been occupied by brimming ice streams since at least the sixteenth century. Gilbert stated in 1899 that "the condition of extreme glaciation . . . does not belong to the series of modern changes. . . . Were it of comparatively recent date the fiord would now be destitute of trees, but such is not the fact. It is true that the slopes are bare in the immediate vicinity of the glaciers, and that the valley walls enclosing the greater glaciers . . . support no trees, but the lower parts of the fiord walls are elsewhere covered by a hemlock forest.

As to the proper interpretation of the peculiarities of forest distribution the case is not altogether clear. In other localities there has seemed good reason to ascribe absence of forest to recent occupation by ice, but here there is a sort of transition from forest to barren which suggests climatic limitation. In the zone of transition the trees are not young and vigorous, as when invading newly-acquired territory, but scrawny and ill-favored, as though struggling desperately against the attack of hostile conditions. In an illustration, representing the edge of the forest nearest to Harriman Glacier, the rareness of branches on the side toward the water suggests that winter fogs driven landward overwhelm the boughs with loads of ice."

From this and from our own observations we are convinced that the presence or absence of trees and their size are not absolute criteria for use indiscriminately in interpreting glacial history where human records are lacking. That the rate of tree growth varies greatly under climatic, soil, and drainage conditions in this region is shown by the great variations in the relation between diameter and number of annual rings already quoted, and by the following record of a spruce found near Baker Glacier in 1910. The last ten years of growth were very rapid, as shown by the annual rings, the previous ten were very slow indeed, and the first 75 years were rapid. With evidence of decades of

[1] Muir, John, Harriman Alaska Expedition, Vol. I, 1902, p. 135.

[2] Fernow, B. E., Harriman Alaska Expedition, Vol. II, 1902, p. 255.

retardation such as this, it is, of course, not safe to estimate the ages of trees by diameter alone. The trees must be cut down and the rings counted.

The absence of trees in southern Harriman Fiord, bringing up the possibility of recent expansion of Harriman Glacier, may be due to the fact that conifers are not always seeded at once after a glacier retreats, while alders and willows may start sooner. Accordingly the length of time since retreating glaciers have uncovered a tract, as determined by vegetation, should always be regarded as a minimum. In the case of the barren zone around Barry Glacier, however, we differ from Grant and Higgins who believe that the advance to its outer limit was probably 25 years or more ago.[1] We have direct evidence that seeding did take place there at once, for in the field we first determined, by ages of trees that the last advance was at least 11 years before (1899 or earlier), and subsequently after reaching home, found the records of the army officers who witnessed the activity in May, 1898, exactly 12 years before. This advance ceased during that year, however, and there was considerable melting before June, 1899, so that willows and alders were able to start on the abandoned lateral moraines and in the edges of the barren zone during the summer of 1899. Thus they were eleven years old in 1910. Similar immediate occupation by vegetation was found to have occurred on the elevated beaches in Yakutat Bay, where, as in the Barry Glacier barren zone, the proximity of mature plants gave opportunity for immediate seeding.

With these facts and limitations in mind, the following relationships of vegetation to the Barry, Serpentine, Toboggan, and Baker Glaciers in Harriman Fiord may be profitably considered.

As already stated, there are conspicuous barren zones around the termini and lower borders of a number of the Prince William Sound glaciers (Pls. CXXXIX and CXXXII, B), as a result of the fact that the glaciers have melted back from their farthest recent advance. There was such a barren zone around Columbia Glacier when Gilbert visited it in 1899, but before 1909 most of this barren zone was overridden by the advancing glacier and by 1910 it was covered, as were most of the barren zones which had bordered the College Fiord glaciers from 1899 to 1909. In Harriman Fiord, on the other hand, these barren zones were still present near most of the glaciers in 1910.

Our first impression was that the advances which caused these barren zones were synchronous and that an examination of the ages of shrubs that had sprung up in the barren zones would reveal the approximate date at which the last advance of all the glaciers had taken place.

The vegetation near Serpentine Glacier was studied and in its barren zone none of the scattered willows, alders, and spruces were found to be more than 27 years old. An earlier moraine outside this belt had trees up to 93 years of age. Baker Glacier, just south of Serpentine, had a nearly-barren inner moraine with shrubs 18 years old and a more-thickly-forested outer moraine with spruces up to 110 years old. Toboggan Glacier had 20 year old shrubs on the last moraine of its barren zone and conifers of 70 years growth on an older moraine. Barry Glacier was bordered by a barren zone with the oldest shrubs only 11 years old; and here there are not two belts, but one, outside of which, on each margin of the glacier, a very perfect push-moraine ridge separates the barren zone from the thick forest 100 to 225 years old. We cut down many trees and counted rings in determining these ages, only the oldest ones being cited here. At the

[1] Bull. Amer. Geog. Soc., Vol. XLIII, 1911, p. 331.

borders of these older deposits there are usually moraine ridges, beyond which is thick-set, mature forest with trees from 100 to 325 years or more of age.

It was found, therefore, that the advances were not synchronous, and that in three cases out of the four the glacier had advanced twice, for it is hardly conceivable that each glacier stood at the inner moraine long enough for the shrubs to grow for 50, 66, and 92 years. The data for the last of these two advances may be arranged as follows, suggesting glacial maxima which terminated in the years shown in the right-hand column, or at some preceding date, but never later.

Glacier	*Age of Trees*	*Earliest Possible Date of Advance*
Serpentine	27 years	1882
Toboggan	20 "	1889
Baker	18 "	1891
Barry	11 "	1898

There are no known visits of white men by which to check any of these dates except in the case of the Barry Glacier, which had retreated slightly from the edge of the barren zone when the Harriman Expedition visited it in 1899. In 1898 some army officers saw it from a distance of ½ mile and the advance was evidently still taking place, for an abnormal number of very large icebergs was being discharged.

Was this group of advances due to climatic variations or to earthquake avalanching? If it was the former we must explain why all these glaciers, which are fed from adjoining snowfields, did not advance approximately together, instead of through a period of 16 years. If earthquake avalanching is thought to have caused the advances we see at once the likeness in lag due to size, observed in the Yakutat Bay advances, for Barry Glacier is longer than Serpentine, and Baker is longer than Toboggan Glacier. The two last are shorter than Serpentine, but have very different grades and shapes of snowfields.

There was a severe earthquake in this part of Alaska in 1880, and the advances show an apparent time relationship to this that warrants the retention of the earthquake avalanching hypothesis as certainly possible, the first glacier advancing two years after the earthquake. As in the Yakutat Bay cases, the last glacier advanced and ceased its advance during a single year, as possibly the earlier ones did also.

In the case of the group of earlier advances the same choice of causes is possible. The short Baker Glacier advanced before the long Serpentine, with the Toboggan advancing still later. The absence of a record for Barry Glacier suggests that the 1898 maximum went beyond this earlier one and destroyed the records of it, as Columbia Glacier has now absolutely destroyed all the evidence of the 1892 maximum, which was plainly visible from 1899 to 1909. The group of advances in Harriman Fiord 70 to 110 years ago, with progressive development of advances of adjacent glaciers of several sizes suggests the possibility of an origin by earthquake avalanching, as the synchronous advance of the eight, variable-sized, contiguous, College Fiord glaciers in 1910, described in the preceding chapter, strongly suggests the probability of climatic origin.

It is to be hoped that some future expedition to Harriman Fiord may determine whether the Harriman, Roaring, Cataract, and Baker Glaciers, which were advancing slightly in July and August, 1910, continued their activity after that year; and whether the Barry, Serpentine, Detached, Surprise, Dirty, Wedge, Toboggan, and smaller glaciers, which were then retreating, also advanced later in 1910. The relationships of vegetation in the barren zones will be important in determining this.

CHAPTER XVII

THE GLACIERS OF PASSAGE CANAL AND BLACKSTONE BAY

General Description

Passage Canal is a fiord, extending in a general east and west direction in the northwestern part of Prince William Sound. Portage Bay, Blackstone Bay, Cochrane Bay, Port Wells, and Culross Passage are the principal indentations leading out of it. It is about 25 miles long, and from 1 to 2½ miles wide. Esther and Culross Islands flank the mouth of Passage Canal and on the western side of Culross Island is Culross Passage, a narrow, crooked channel, 11 miles long, connecting Passage Canal with Port Nellie Juan to the south. These fiords have steeply-rising walls, a great many snowfields near by, and moderate-sized glaciers, some of which extend down to sea level.

Previous Observations

The exploration of these fiords was accomplished by several of the same men who did the early work in the other parts of Prince William Sound, with the exception of the Harriman Expedition which did not visit them. Maps of one kind or another and brief descriptions have been published by Vancouver and Whidbey,[1] Petroff,[2] Applegate,[3] Glenn,[4] Learnard, Castner and Kelly, Mendenhall,[5] Herron,[6] Porter,[7] and Grant and Higgins.[8] Of these observers only Grant and Higgins have given especial attention to glacier study.

The junior author, in charge of the National Geographic Society's expedition of 1910, devoted four days (August 2 to 5) to a study of the glaciers of inner Passage Canal and Blackstone Bay.[9] During this time we made the map of Tebenkof Glacier, the

[1] Vancouver, George, Voyage of Discovery to the North Pacific Ocean and Round the World, Vol. V, 1801, pp. 249–250, 288, 306–311. Map reproduced by Davidson (see below) Map V.

[2] Petroff, Ivan, Tenth Census of the United States, 1880, Vol. VIII, 1884, p. 27 and Map VI, facing p. 75.

[3] Applegate, S., map in Davidson's Glaciers of Alaska that are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc., Pacific, Vol. 3, 1904, p. 27 and Map XI.

[4] Glenn, E. F., War Dept., Adj.-Gen. Office, No. XXV, 1899, pp. 17–18, 26–32; Learnard, H. G., *Ibid.*, pp. 125, 127–8; Castner, J. C., *Ibid.*, pp. 189–192; Kelly, L. S., *Ibid.*, p. 289, also map in pocket.

[5] Mendenhall, W. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 273–4, 301, 325–326, and Map 16.

[6] Herron, J. S., War Dept., Adj.-Gen. Office, No. XXXI, 1901, p. 13 and map.

[7] Porter, R. W., Report on Population and Resources of Alaska, Eleventh Census, 1890, Washington, 1893, p. 70.

[8] Grant, U. S. and Higgins, D. F., Glaciers of the West Coast of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLIII, 1911, pp. 402–407; in Reid's Variations of Glaciers, Journ. Geol., Vol. XVII, 1909, p. 671; also maps in Bull. 284, U. S. Geol. Survey, 1906, Fig. 4, p. 79; *Ibid.*, Bull. 379, 1909, Pl. IV, facing p. 88; *Ibid.*, Bull. 443, 1910, Pl. I, facing p. 10 and Pl. II, in pocket.

[9] Martin, Lawrence, The National Geographic Society Researches in Alaska, Nat. Geog. Mag., Vol. XXII, 1911, p. 559; Journ. Geol., Vol. XIX, 1911, p. 458.

modified map of the ice tongues at the head of Blackstone Bay, and soundings throughout Blackstone and Portage Bays.

Blackstone Bay

Topography. Blackstone Bay, a southwestern branch of Passage Canal just west of Port Wells, is about 11½ miles long and from one to two miles wide. It is a steep-walled fiord, but the walls are lower than those of Harriman Fiord and inner Passage Canal which penetrate the very heart of the mountains. Near the head of Blackstone Bay is the long, narrow Willard Island and a projecting peninsula; and around the head of the bay are eight small glaciers. On the southern side of the bay, close to Passage Canal, is a cove about a mile long, a short distance south of which is Tebenkof Glacier, the largest ice tongue of Blackstone Bay.

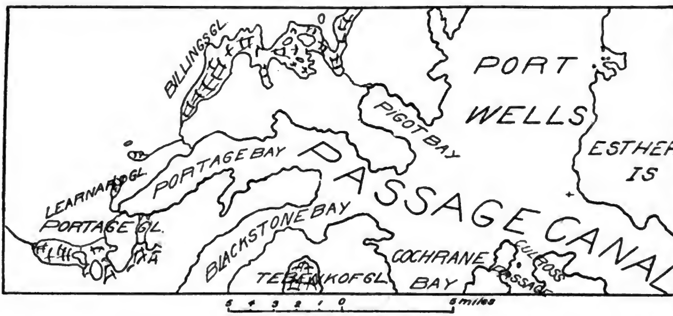
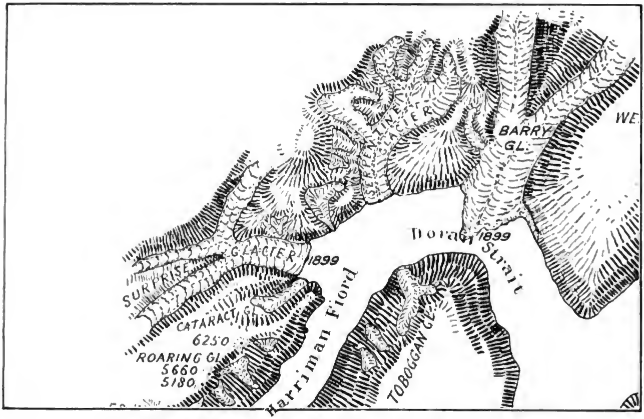


Fig. 51. Sketch of Passage Canal.

Tebenkof Glacier—Snowfields and Ice Tongue. Tebenkof Glacier (Fig. 52) is known to be over ten miles long and is from less than one to over two miles wide in the lower portion. It terminates over a half mile from the head of the cove, into which a number of large streams flow. Tebenkof Glacier also has two subordinate termini, which are lobes projecting from the side of the glacier into tributary valleys. The largest of these is about 4 miles from the terminus and projects westward for about a mile and a half, sending a glacial stream down to Blackstone Bay. The smaller lobe, which is on the eastern side about two miles south of the terminus, projects a shorter distance from the side of the glacier and its stream discharges eastward to Cochrane Bay.

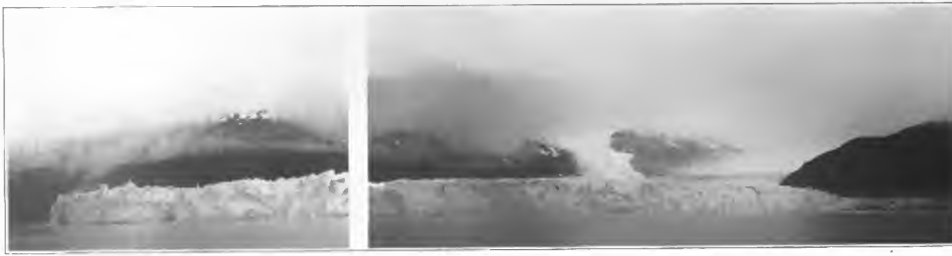
The glacier heads on a high snow divide which is unexplored and it seems probable that the snowfields there coalesce with those that feed the smaller ice tongues to the west at the head of Blackstone Bay and the glaciers on the northwestern side of Port Nellie Juan. At least 10 miles south of the terminus of Tebenkof Glacier, there is another small glacier flowing northeastward from this same snowfield. It is about 2 miles long and its outlet stream flows eastward to Cochrane Bay about 2 miles distant. The absence of glaciers in Cochrane Bay on the east, and in Blackstone Bay on the



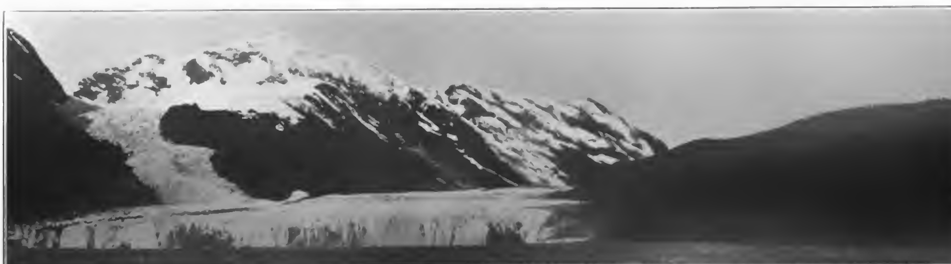
A. SKETCH OF BARRY GLACIER, AS OBSERVED IN 1899 (GANNETT)

B. HARRIMAN FIORD IN 1910
Showing the retreat of Barry and Surprise Glaciers since 1899.

Plate CXXX



A.



B. Barry Glacier in 1899 and 1905, Showing Retreat

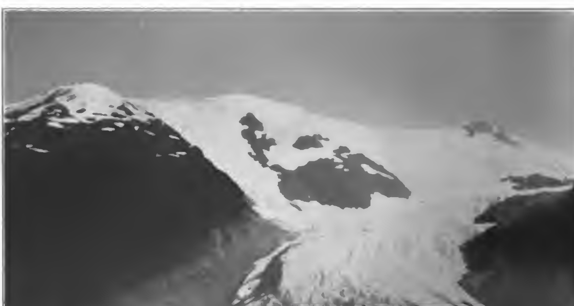
Upper photograph by C. Hart Merriam, from a boat close to Photo Station A. Lower photograph by U. S. Grant and Sidney Paige, from Photo Station A, on Point Doran. Compare with Plate CXXXI.

A.

B. Barry Glacier in 1909 and 1910, Showing Retreat

Upper photograph by U. S. Grant and D. F. Higgins. Lower photograph by National Geographic Society Expedition. Both from Photo Station A, on Point Doran. Compare with Plate CXXX

A. WESTERN LATERAL MORAINE OF FORMERLY EXPANDED BARRY GLACIER
Forming the boundary between the barren zone and the thick mature forest, some of whose trees overturned during an ice advance as seen on the left. Photograph, July 26, 1910.



B. TOBOGGAN GLACIER
Showing also barren zone and outwash plain. Photograph by U. S. Grant and Sidney Paige, August 20, 1905.

Plate CXXXIII

Serpentine Glacier
Rising in cirques at the base of Mt. Gilbert. Curving lateral moraine on right. Photograph from Station DD, July 26, 1910.

PLATE CXXXIV

SURPRISE GLACIER
With Detached Glacier on right. Photograph from Station F, July 27, 1910.

Surprise and Cataract (on left) Glaciers

In 1899 a dirt-stained lateral moraine extending slightly beyond the clear ice of Surprise Glacier reached as far out as Cataract Glacier. Hence a retreat of 1¼ miles has occurred here between 1899 and 1910. Photograph from Station E, July 27, 1910.

PLATE CXXXVI

HARRIMAN GLACIER

Showing snow-covered névé-sheathed slopes. Photograph (copyrighted) by E. C. Curtis, June, 1899. From near Photo Station F.

Western Edge of Harriman Glacier

Photograph (copyrighted) by E. C. Curtis, from Station GG, June, 1899. Ink line shows advance and thickening between 1909 and 1910.

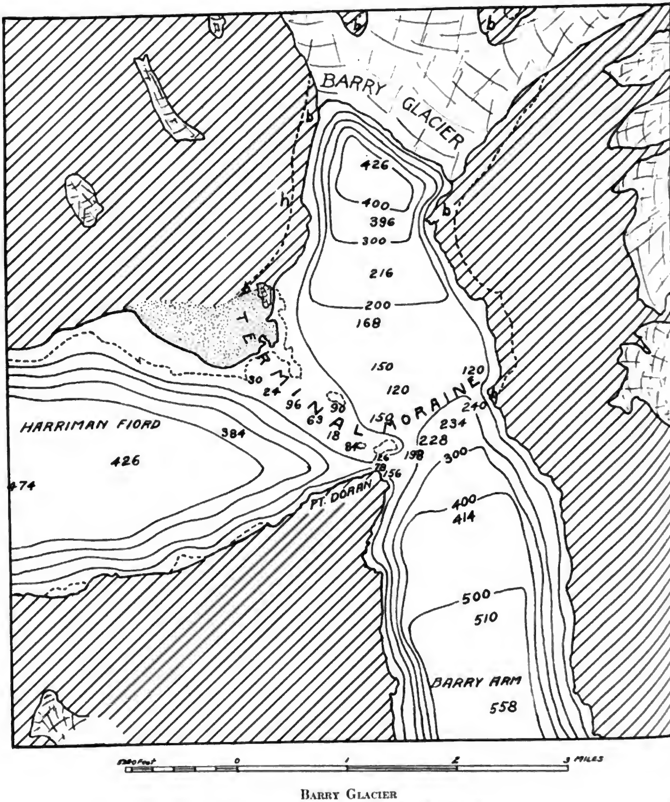
Plate CXXXVIII

A Great Glacial Groove
Cut in hard rock, Point Doran, Harriman Fiord.

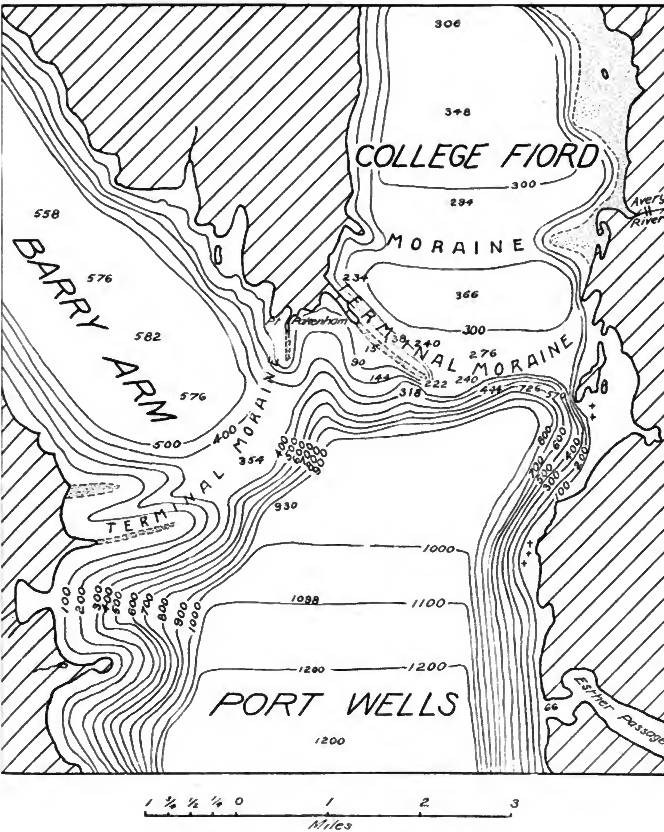
PLATE CXXXIX

WESTERN WALL OF FIORD NEAR BARRY GLACIER

Showing barren zone terminated above by thick forest. Stream emerges from hanging valley with lip 1050 feet above sea level. Successive positions of glacier indicated by ink lines, while in 1898 the glacier rose to the upper edge of the barren zone, and in 1899 there was only a low water fall. Photograph by U. S. Grant and D. F. Higgins, from Station C, June 29, 1909.



Barry Glacier
Showing crescentic submarine terminal moraine.



Moraine Bars at Head of Port Wells

Plate CXLII

Terenkof Glacier in 1910 from Station B

Plate CXLIII

Eastward-Descending Rock Benches

Due to glacial erosion, on northern side of Passage Canal near Portage Glacier. Photograph by W. C. Mendenhall, 1898.

PLATE CXLIV

THE THROUGH VALLEY OF PORTAGE GLACIER

As seen from Turnagain Arm. This ice tongue was used in 1898 as a highway across Whidbey Isthmus between Prince William Sound and Cook Inlet. Photograph by F. H. Moffit, August, 1904.

west, everywhere excepting at the head, emphasizes the elevated position of Tebenkof Glacier which lies between the two bays and which extends nearly as far north as their mouths.

The surface of the glacier, which has no medial or lateral moraines, is exceptionally

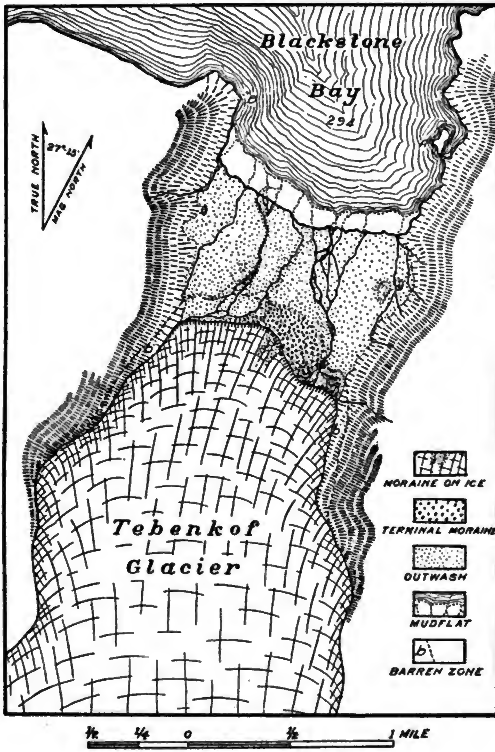


FIG. 52. TEBENKOF GLACIER IN 1910.

clean (Pl. CXLII) and without morainic débris except in little patches at the terminus. It was so slightly crevassed in August, 1910, that travel over the glacier surface was very easy. The surface slopes 18° to 26°, steepening to 32° and 40° near the terminus

and in one place, near the western edge, to 45°. It is separated from the sea by a crescentic terminal moraine and an outwash gravel plain.

Lack of Change. The terminus of the Tebenkof Glacier in 1910 (Fig. 52) was not very different from the conditions in 1909 and, so far as information is available, in previous years. Applegate showed the glacier in 1887 in about the same position as in 1909 and 1910. Photographs from sites occupied by Grant in 1909 show no change during the following year on the eastern edge and only a slight retreat at the western margin of the ice front. There are no earlier photographs.

Terminal Moraines. The terminal moraine is close to the glacier and varies considerably in character. The portion near the eastern center (Fig. 52), having a width of from 700 to 1150 feet, has a hummocky inner surface against which the edge of the ice rests. On it are a few pools, but no large kettles. There is a similar small area near the eastern valley wall. The remainder of the terminal moraine forms a striking contrast with this, being a single ridge or series of small parallel ridges less than 100 feet wide, 20 feet or less in height, and extending across the valley with a shape similar to that of the present terminus of the glacier and at a distance of 600 to 800 feet from it.

Between this terminal moraine ridge and the glacier, both east and west of the broader central portion of the moraine, outwash gravel is still being deposited, the gravel deposit on the east containing small pools possibly due to slumping of a small buried portion of the glacier front. As these gravels were still being built up by glacial streams in 1910, it seems probable that other parts of the terminal moraine deposits have been cut away by these streams and buried by their accumulations.

On the eastern side of the moraine, which was barren in most places, were scattered alders up to eight years old in 1910. On the western side, the outer part of the moraine had willows 10 years old and alders of 12 years growth.

This terminal moraine is not continued by lateral moraines at the borders of the glacier, probably because little morainic material was supplied from the bare rock there. There are, however, discontinuous, linear patches of moraine paralleling the ice edge on the eastern side.

Older Terminal Moraine. On the western side of the valley and about 50 feet outside the terminal moraine, was a fragment of an older terminal moraine, contrasting with the inner one in having much moss and many shrubs growing upon its surface. One willow examined was 18 years old. This furnishes the only evidence of two stages in the recent history of Tebenkof Glacier.

Outwash Plain. The outwash gravel plain, with a length of 2800 to 4500 feet and a width of about a mile, is, in some parts, still being built up by streams. To the northward it is continued into the bay by a mud flat 400 to 700 feet wide at low tide. This plain is 35 to 50 feet above sea level at the ice edge and slopes at the rate of 60 to 65 feet per mile. Some parts of the outwash gravel plain are, however, no longer being built up and have had no streams for over 20 years, as is shown by an extensive growth of willow and alder thickets. Less than half way between the bay and the glacier are several clumps of tall, symmetrical spruce trees, the largest being about 3 feet in diameter and with an estimated age of a century or more. A pair of small hillocks, which appear from a distance to be rock hills, rise up through the alluvial deposits on the eastern edge of the outwash gravel plain; and there is a single, similar hillock on the western side near the valley wall.

Barren Zone. On each side of the terminus of Tebenkof Glacier was a barren zone (bb, Fig. 52), also noted in 1909 by Grant and Higgins. It is 200 to 400 feet in width, but wedges out at a distance of less than a mile to the southward. The upper edge of this barren zone marks the greatest expansion of Tebenkof Glacier in recent years, for above it is thick mature forest, with trees 16 to 24 inches in diameter and with a few dead trees, still standing at the edge of the forest. The western margin had many dead tree heaps at the edge of the barren zone.

Between the barren zone and the ice the rock surface is prevailingly bare, and upon it are scattered fragments of dead tree trunks, while, where there is soil, flowers, grasses, and scattered young willows, alders and spruces are growing. The upper part of the western barren zone has quite a number of scattered shrubs, but its lower part has almost none. The largest of the alders observed had seven annual rings. This and the 12 year old alders on the terminal moraine indicate that the last expansion of the glacier up to the edge of the barren zone and out to the terminal moraine was at least 12 years before 1910.

Glaciers at the Head of Blackstone Bay—Relationships of Glaciers. This is a group of moderate-sized ice tongues, descending from névé fields which surround the head of Blackstone Bay. All of them are cascading glaciers of varying steepness, and only two are tidal, though three others on the eastern side reach almost to tidewater. The name Blackstone Glacier has heretofore been used for the whole group, as if they were distributaries of a single ice cap, but, as they are not lobes of a single glacier, but separate valley glaciers, as few of their névé fields coalesce on the ridges between, it seems desirable to give separate names to each of the ice tongues. The additional names are taken from colleges in Wisconsin. The apparent coalescing of the glaciers, indicated in Fig. 53, would practically disappear if more detailed maps were made which separated the ice tongues from the névé on the adjacent mountain slopes. Ripon Glacier, for example, heads in a cirque which has névé all over its head and side walls.

At a period of greater expansion all of these glaciers were tributaries to a great ice tongue in Blackstone Bay similar in dimensions to the Tebenkof Glacier. It seems probable that the retreat of this glacier and its separation into distinct ice tongues, while the non-tidal Tebenkof Glacier still extends over 6 miles farther northward, is, in part at least, a result of more rapid melting of the Blackstone ice tongue on account of its ending in salt water.

Blackstone Glacier. Blackstone Glacier heads in unexplored snowfields southwest of the bay, having a moderately-sloping upper portion, fed by numerous small tributaries, and a steeply-cascading terminus, which descends several hundred feet in the last quarter of a mile, terminating in the southwest corner of the bay in a vertical ice cliff over a hundred feet high, from which small bergs are discharged. The glacier is a clean white, severely-crevassed mass reaching tidewater only in the eastern portion, but descending nearly to tidewater for some distance to the west.

On the early maps by Applegate, Mendenhall, and others this glacier is indicated as coalescing with the adjacent Beloit and Northland ice tongues and as extending over half a mile farther north to the southern end of Willard Island. This is probably an error, for Applegate did not go near enough to the glacier to be sure of its relationships, and in 1910 spruce trees, 10 to 12 inches in diameter, were found at sea level on the

southern end of Willard Island. The correct position of the terminus of the glacier was first mapped in 1909 by Grant and Higgins.

A comparison of their photographs with ours taken in 1910 shows a slight advance of the glacier during that year. It is not certain exactly how much advance there was, but a rock ledge exposed in 1909 beneath the eastern side of the terminus was overridden in 1910 and the tidal frontage was slightly increased. The western edge, toward Northland Glacier, covered more of the bare rock slopes near sea level in August, 1910, than in July, 1909. At several points along this western portion of the terminus, ice blocks were sliding down from the end of the glacier into the sea as a result of this advance.

Beloit and Marquette Glaciers. These are two adjacent cascading glaciers, Beloit Glacier being at the very head of Blackstone Bay, while Marquette Glacier lies a short distance northeast of it. The latter has a much gentler slope than the former. Beloit Glacier descends from a gently sloping upper portion, at about the same level as Blackstone Glacier, and terminates at sea level in a vertical ice cliff from which bergs were discharged in 1909 and 1910, there being no appreciable change in the ice front during the period of observation. Both of the glaciers are without medial or lateral moraines, Beloit Glacier being severely crevassed, while Marquette Glacier is rather smooth. The former is fed from névé fields adjoining the Blackstone Glacier and is separated from the Marquette Glacier by a mountain ridge, the upper portion of which is covered with névé which feeds into each of the glaciers. In 1909 and 1910 Marquette Glacier terminated a few hundred feet from the water's edge and underwent no appreciable change during that period. It was covered with snow nearly to sea level in August, 1910.

Ripon and Lawrence Glaciers. These two cascading glaciers are on the eastern side of Blackstone Bay, Lawrence Glacier being about three quarters of a mile northeast of Marquette Glacier, and Ripon about the same distance farther to the northeast. On Applegate's map both of these glaciers seem to reach tidewater, but the author of the map did not go near enough to them for us to feel certain that this is correct. In 1909 and 1910, each of the glaciers terminated 150 to 200 feet from the fiord.

Each of the glaciers is moderately crevassed and without moraines. They are both deeply set in mountain valleys, separated by broad rock spurs and are fed from extensive névé fields which mantle the whole mountain slope. Their surface slope is 18° to 20°, or about the same as Marquette Glacier, and not nearly so steep as the slope of the Beloit and Blackstone Glaciers. The névé fields on the slopes of Mt. Applegate, east of Ripon Glacier, mantle large areas and feed an unnamed glacier lobe in the next valley to the northeast, as well as a number of other small lobes.

In 1910 portions of the terminus of Lawrence Glacier were so near the water's edge that ice fragments were sliding down the bare rock slope into the bay, so that this glacier was discharging small icebergs without being a tidal glacier. A small, but clear-cut, stony, push moraine, forming the boundary of a moderate-sized barren zone between the moraine and the ice, extended around parts of the terminus of Lawrence Glacier, excepting in the middle where the ice fragments were discharged. Within this barren zone there were only scattered young shrubs, while dense, mature forest grew outside.

The terminus of Ripon Glacier lies farther back from the sea than that of Lawrence Glacier, ending on bare rock ledges. The nose of the glacier was mantled in places

with ablation moraine, which gives part of it a fictitious appearance of being lobate. A crescentic terminal moraine marks the site of a former extension of the glacier and inside of it is a barren zone, the largest in Blackstone Bay. It includes a large area of bare rock with scattered young shrubs, and outside of it is the normal mature forest.

There is a possibility that this glacier has had a double history of recent advances, for the part of the barren zone nearest the glacier had almost no vegetation and was separated by a small, low moraine from an outer portion which had a great many more young shrubs. The age of this youngest vegetation near Ripon and Lawrence Glaciers was not determined carefully, but may be estimated as certainly less than 20 years.

Glaciers on the Western Side of Blackstone Bay. The Northland, Milton, Concordia, and Carroll Glaciers are fed from névé fields west of the Blackstone Glacier. None of them are as large as the glaciers on the eastern side of the bay, and all excepting the Northland Glacier terminate from a quarter to a half mile from the bay. Northland Glacier extends down nearly to the branch of the bay west of Willard Island.

All of these glaciers are without moraines, and are more or less crevassed, the Carroll, Concordia and Milton Glaciers, which are descending steeper slopes than the Northland Glacier, being most crevassed. There is not enough previous information about these glaciers to determine their recent history, although the sliding down of fragments from some of the termini, and the absence of barren zones, suggested the possibility of recent advance.

Glacial Erosion—Erosion Above Sea Level. Evidence of glacial erosion in Blackstone Bay is found in the form of the fiord above sea level, and especially in the steepened walls and bare rock slopes scored with glacial striae. Willard Island shows this glaciation particularly well, having numerous typical roches moutonnée forms, and good-sized rock basins containing lakes. The northern wall of Blackstone Bay at the curve, eight miles north of Blackstone Glacier, is far steeper than the opposite wall, suggesting glacial undercutting on the outer side of this curve. There are also hanging valleys illustrated by the valleys of Carroll and adjacent glaciers, and by a valley on the eastern side of the fiord about three miles north of Ripon Glacier. Its lip is a few hundred feet about sea level, and it has no glacier at present.

Submarine Erosion. The portion of the fiord below sea level, as revealed by soundings, also emphasizes the great amount of glacial erosion. The fiord bottom slopes at the average rate of about 75 feet to the mile, having a depth of 216 feet near the Blackstone and Beloit Glaciers, and 1080 feet where Blackstone Bay enters Passage Canal 11½ miles distant. Half way down the fiord, however, there is a deeper point (1188 feet); and there is a steep descent of over 500 feet, 2½ miles from the head, besides other interruptions of slope, described below. The visibly steepened slopes of the side walls of the fiord seem to continue below sea level, suggesting that it has the normal relatively flat-bottomed cross section.

The slope of the bottom of this fiord from Beloit Glacier to the southern end of Willard Island is at the rate of about 100 feet to the mile. Then comes a conspicuous increase in the slope with a descent of 462 to 522 feet in less than three-quarters of a mile, or at the rate of over 650 feet to the mile. Such a confluence step in the fiord bottom is a feature often observed in glaciated mountain valleys and the same condition exists above sea level at the southern end of Blackstone Bay where Blackstone and Beloit Glaciers descend a similar step of even greater height. A possible explanation of this

step is found in the relationship of the channels on the east and west sides of Willard Island. It is possible that when the glaciers from either side of Willard Island united, the combined ice tongue had such increased power as to enable it to erode more deeply beyond the end of the island. Northward to the mouth of Blackstone Bay the fiord bottom slope averages 30 to 85 feet to the mile.

The channel bottom between Badger Point and the southern end of Willard Island is basin shaped, with only 90 feet of water at the northern margin and 126 feet in the center. This may be a rock basin due to glacial erosion. There may also be a rock basin southwest of the point where there is a depth of 936 feet, in Blackstone Bay, near the turn just west of the cove of Tebenkof Glacier. The increased depths in the fiord north of Tebenkof Glacier may be due to increased glacial erosion resulting from the union of this glácier with the Blackstone Bay ice tongue.

The relationship of the Tebenkof Glacier cove to Blackstone Bay is that of a submerged hanging valley, the lip of the hanging valley at the bottom of the cove being at least 700 feet above the floor of Blackstone Bay.

Glacial Deposits—Above Sea Level. As is the case in the other fiords of the Prince William Sound region, glacial deposits above sea level are relatively unimportant. The largest accumulation of this kind is the terminal moraine and outwash plain near the terminus of Tebenkof Glacier, already mentioned. There are also minute terminal moraine deposits around the borders of the Ripon and Lawrence Glaciers, and there is a thin mantle of ground moraine on the fiord walls, where not too steep, and on parts of Willard Island.

Moraine Bar at Willard Island. The only other morainic accumulation in the bay is of considerable interest. It was sketched by Grant and Higgins in 1909, who referred to it as "modified remains of two recessional moraines." This partly-submerged terminal moraine, or moraine bar, nearly joins Willard Island with the mainland on the east (Fig. 53). It projects over three-eighths of a mile southeastward from Willard Island, extending more than half way across the fiord; further east is allow. islet; and then comes a point which projects a quarter of a mile from the main and At low tide nearly the whole of the moraine bar is exposed, there being only two narrow channels across it, the easternmost only a few feet deep, the westernmost, 24 feet deep at about mid-tide. As the submerged contours show, the fiord bottom here rises about 474 feet above the bottom of the bay immediately to the north and 210 to 288 feet above the bed of the fiord to the south. The northern face of the moraine, therefore, slopes at the rate of 1900 feet to the mile, while southward toward the glacier, the slope is only 1120 feet per mile.

Some beach sand is found on portions of this moraine and on those parts of the bar that rise entirely above high tide trees and grass are growing. That this bar is not a mere sand spit is shown conclusively by the character of the foundation material which is especially well exposed on the portions uncovered at low tide. It consists of angular bowlders up to 6 feet in diameter, some of them striated, which could have reached their present position only in the glacier, being too large and too numerous for an iceberg deposit with sea level as at present. The till at the surface of this terminal moraine has largely been washed away by the tidal currents, which at high tide sweep strongly across the submerged parts of the bar.

It is clear that a moraine exists here on the surface. Whether it is (1) a thick till

deposit,—a moraine rising 210 to 474 feet above the fiord bottom and all submerged excepting the top,—or (2) a thin superficial deposit, resting upon a rock swell, cannot be determined from the facts at hand. The lack of thick corresponding deposits upon

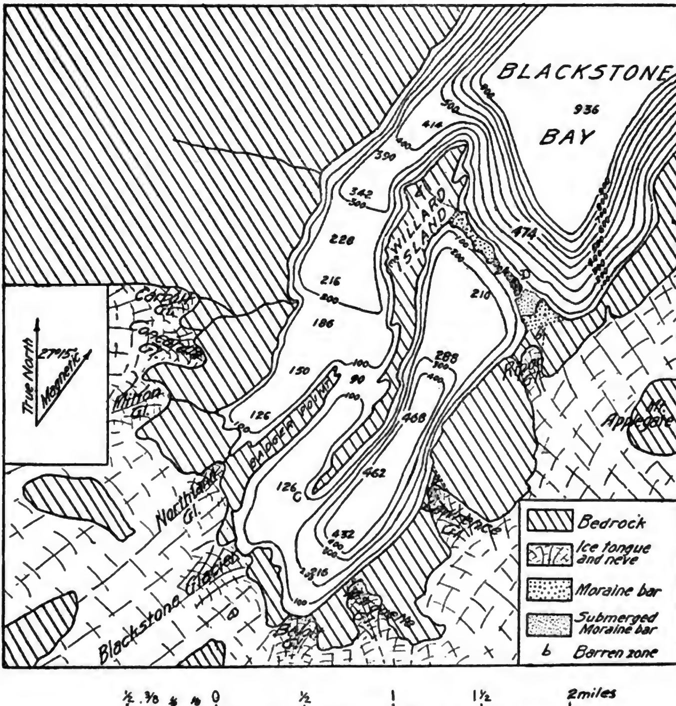


FIG. 53. GLACIERS AND SUBMARINE CONTOURS OF BLACKSTONE BAY. Soundings in feet. Photographic stations shown by letters C, D, and E.

Willard Island and in the fiord to the west of the island suggest the latter. The broadening of the eastern passage just here also accords with this view, for glacial erosion would naturally be efficient in the constriction to the south (462 to 468 feet deep), and the incoming of the steeply-descending Ripon Glacier might have caused a plunging of the ice and an increased efficiency of erosion to the north where the depths are 474 to 936 feet.

On the other hand, the Ripon Glacier is not in exactly the right position for this, and the narrowness of the shoal area is difficult to explain by glacial erosion and easy to understand if the whole deposit is a submerged moraine. The coincidence of a halt of the glacier exactly on an uneroded rock swell is also difficult to explain, if the moraine is thought of as resting so precisely on an irregularity in rock. There is no special rock texture or structure to help explain this phenomenon, if mainly a rock feature. The geological map by Grant and Higgins [1] indicates that Willard Island and the whole shore line of Blackstone Bay are made up of graywacke and slate and small dikes with no observed peculiarities, such as would fit in with the idea of differential erosion due to the nature of the rock. On the whole, therefore, we believe that the facts point toward a moraine bar here, rather than a narrow upward swell in the bedrock floor of the fiord.

Relationships of Forest to Stages of Glaciation. As noted by Grant and Higgins, the southern six-sevenths of Willard Island constitutes a relatively barren zone, though not one of the extremely young barren zones like those near Ripon and Lawrence Glaciers, or the ones near Barry and other glaciers in Harriman Fiord, for within it there are a few trees up to 10 or 12 inches in diameter. Throughout this somewhat overgrown barren zone, there is a scattered growth of evergreen trees and many alder thickets, but the larger part of the area has no trees, though in the open spaces thickly covered with grasses, moss, and peat. On the north it terminates along a definite line (marked as the limit of the barren zone, b, Fig. 53), extending westward across Willard Island from the base of the morainic bar. North of this line there is thick, mature forest, and it therefore seems clear that an advance of the former glacier of Blackstone Bay extended down to this line some scores of years and possibly over a century ago. The fact that the edge of this barren zone continues the line of the moraine bar just described, suggests the association of the moraine with this advance. It is to be noted, however, that we found no similar moraine at the western end of this zone, in the channel west of Willard Island; though the fact that our soundings there were a quarter of a mile apart, makes it possible that there is a narrow submerged moraine which we did not discover.

All about lower Tebenkof Glacier and throughout the northern end of Blackstone Bay, there is thick, mature spruce forest extending from timber line to sea level. The forest ends about a mile north of Carroll Glacier on the western side of the bay and at Ripon Glacier on the eastern side; but there is scattered vegetation beyond this. Badger Point, for example, has scattered trees and thick groves with individuals apparently as old as those on the southern part of Willard Island; and at moderate heights above sea level there are a few clumps of good-sized spruces even on the mountain spur between Beloit and Blackstone Glaciers. The trees nearest the glaciers, as on Badger Point and between Beloit and Blackstone Glaciers and near Tebenkof Glacier, are tall and well formed, but some of the trees in the southern part of Willard Island are much stunted, very thick in proportion to height, and with limbs only on the northern side, showing that in exposed places the glacier wind, here from the south, has retarded tree development. Even with the limitations, resembling the conditions in southern Harriman Fiord, which of course lengthen the time since great glacial expansion, the abrupt limit of mature, thick forest on Willard Island, coinciding, as it does,

[1] Bull. 443, U. S. Geol. Survey, 1910, Pl. II.

with the position and trend of the terminal moraine to the east of this island, has been regarded, both by Grant and Higgins and by the junior author, not as a feature related either to climate or soil, but as clear proof of a maximum of the Blackstone, Beloit, and adjacent glaciers a century or two ago. Whether this followed a long period of glacial retreat during which the thick forest advanced to the very head of Blackstone Bay is unknown. This maximum may have been as recently as 116 years ago, for Whidbey's map made in 1794 shows Blackstone Bay much shorter ("two leagues and a half" in length) than in 1910, though it is not detailed enough for us to be sure of the conditions; and, as usual in Vancouver's maps, no glaciers are shown. Willard Island is not indicated at all in the 1794 map.

The barren zones around the termini of Tebenkof, Ripon, Lawrence and other glaciers (bb, Fig. 53), with their youthful shrubs, attest to a more recent period of glacial change. As already pointed out, this was double in the case of Tebenkof Glacier, whose terminal moraines have shrubs 12 and 18 years old respectively, and perhaps double in the case of Ripon Glacier, in whose barren zone there appear to be moraines and shrubs of two ages.

Passage Canal

General Description. Passage Canal, as shown in Fig. 51, is an irregular fiord from which Port Wells, Blackstone Bay, and other branches extend. The outer portion from the entrance of Blackstone Bay and Port Wells eastward to Prince William Sound has no ice tongues whatsoever; but, as we have seen, there are numerous glaciers in the fiords that branch from it. The inner extension of Passage Canal, called Portage Bay (Fig. 51), also has ice tongues, the most conspicuous being two good-sized glaciers on the northern side, and a group of three around the western end (Fig. 54). Since none of these reach tidewater, Portage Bay is free from icebergs, excepting as they occasionally float in from Blackstone Bay or Port Wells.

Seth Glacier. This glacier, which is fed from the snowfield area from which Pigot Glacier flows eastward toward Port Wells, extends southward with a length of about 1½ miles and a width of ½ mile, terminating a little over a mile north of a large cove on the northern side of Passage Canal. The glacier surface is clean, with one medial moraine near the western side, and from the fiord seems to be little crevassed. Its condition previous to 1909, when Grant and Higgins mapped it from a distance, and 1910, when we saw it, is not known in sufficient detail to describe its recent history.

Névé-Sheathed Mountain Slopes. A large cirque east of Seth Glacier, and between 400 and 500 feet above sea level, contains an ice mass; and parts of the mountain slope between this cirque and Seth Glacier contain large névé fields which mantle the mountain slope, making a type of glacier similar to the thin sheath of ice on the slopes around the Harriman Glacier, and between glaciers around the head of Blackstone Bay. A similar névé mass east of Port Wells, and just north of Esther Passage, has been given the descriptive name of Cap Glacier by Grant.

Billings Glacier—The Ice Tongue. This is the largest ice tongue in Passage Canal, having a length of over four miles and a width of about a mile. It is fed from snowfields adjacent to those that supply Seth Glacier on the east and Harriman Glacier on the north and terminates between 1 and 1½ miles from the fiord.

Billings Glacier is fairly clean, with no moraines of any size. Its lower end has a very moderate slope, and its front has apparently been in about the same position and condition in 1887, when it was first shown on a map by Applegate, in 1898, when some of the army officers of Glenn's party spoke of it as a "dead glacier," in 1909 when it was sketched by Grant and Higgins, and in 1910, when it was seen from the fiord by the National Geographic Society's expedition. No one seems to have gone near enough the glacier to study it in detail.

The Valley Train. The largest stream from Billings Glacier, Cabin Creek, flows southwest and south from the terminus of the glacier over an outwash gravel plain or valley train, receiving the stream from a small unnamed glacier on the west, and entering the fiord over a large delta, which has been built forward nearly a quarter of a mile into Passage Canal. The larger part of the outwash plain and delta is overgrown with willows, alders, and cottonwoods, showing not only that the glacier has not extended down to the fiord for many years, but that its melting does not supply as much water and detritus as formerly and that the building of the outwash of the valley train is now going on slowly; otherwise the streams would shift back and forth over the valley bottom and destroy the vegetation.

A Marginal Gorge. Another glacier stream from Billings Glacier enters the fiord a short distance to the northeast (g, Fig. 54) and is of special interest because between the glacier and the fiord it flows through a narrow, steep-sided rock gorge similar to but deeper than the abandoned gorges on the northern side of Toboggan Glacier. The gorge has been cut so nearly down to sea level that, through it, glimpses of the glacier surface may be had from the fiord. The stream seems to have established this course at a time of greater expansion of Billings Glacier when the ice was several hundred feet thicker than now, and extended high enough for a marginal stream to establish the present channel across the rock spur into which it is deeply incised. The glacier must have long remained in this more expanded condition to have permitted this gorge to be cut so nearly down to sea level that it is still occupied by a stream, although the ice tongue long ago retreated far enough to render this channel unnecessary. It is possible that this expanded stage occurred at the time when the glacier at the head of Passage Canal extended eastward, receiving Billings Glacier as a tributary, so that the broad valley now occupied by the main stream on the outwash gravel plain was entirely filled with ice, necessitating the cutting of this marginal channel.

Glaciers at the Head of Passage Canal—Portage Glacier Pass. There are three ice tongues at the western end of Passage Canal, of which the middle one, Portage Glacier, is the largest. It is of particular interest because it occupies a low pass across the Kenai Mountains which was habitually crossed by the Alaskan natives and by the Russians previous to the explorations of Vancouver and Whidbey in 1794, and probably later. The glacier now fills this pass but it has, nevertheless, been used as a highway by a number of United States army parties, and in 1898 by many prospectors. Their route, described by Mendenhall, led three quarters of a mile over an outwash gravel plain to the glacier, up Portage Glacier to the divide, and down the connecting through glacier on the western side, to an outwash gravel plain at a distance of about eight miles from the head of Turnagain Arm, a branch of Cook Inlet. The total length of this portage route is between 12 and 13

miles,[1] though the isthmus is only 10½ miles wide in a straight line. Davidson has proposed that this be called Whidbey Isthmus.[2]

There is also a difficult, roundabout, non-glacial route from the delta of Cabin Creek, on Passage Canal, through a series of valleys to the north of Portage Glacier to Cook Inlet. This route was traversed by army parties in 1898 and 1899. Along it there are several small ice tongues, shown on maps by Kelly and Herron, but not otherwise described.

Portage Glacier. Portage Glacier, as described by Petroff, Mendenhall, Learnard, and Castner, is fed from snowfields that lie to the north and south of the pass. There are three tributaries from the southern side and one or two from the northern, which unite to form an east-west trending glacier between 4 and 5 miles long, ¼ to ½ mile wide, and with the part flowing westward toward Turnagain Arm much the longer. The divide in this saddle-like glacier is about a mile from the eastern terminus at an elevation of between 1000 and 1100 feet, and both divisions of the glacier slope moderately, excepting at the divide, which has a steep slope on the eastern side. Each terminus lies almost at sea level, though some distance back from it.

Portage Glacier is best shown upon the army officer's 1898 map and upon the map made in 1909 by Grant and Higgins (Fig. 51). It is also shown, less in detail, on the 1898 map published by Mendenhall and copied on Hamilton's map,[3] and on a second U. S. Geological Survey map [4] where the eastern terminus of Portage Glacier is shown incorrectly as a tidal glacier ending in Passage Canal. This error has also been followed in a Coast Survey chart.

Advance of Portage Glacier. When Lieutenant Whidbey arrived at the head of Passage Canal on June 7, 1794, Vancouver states that he found that he "had approached within twelve miles in a direction S. 60 W. of the spot where . . . (he) had ended his examination of Turnagain Arm. The intermediate space was an isthmus so frequently alluded to before, on either side of which the country was composed of what appeared to be lofty, barren, impassable mountains, enveloped in perpetual snow; but the isthmus itself was a valley of some breadth, which, though it contained elevated land, was very free from snow, and appeared to be perfectly easy of access; a little to the eastward of this valley, a rapid stream of fresh water [5] rushed down a gully in the lofty mountains, and found its way to the sea through a margin of low land extending from the base of the mountains, and producing pine trees, cranberries and a few other shrubs. On the western point of entrance into this brook was a small house. . . .

This house and the general appearance of the country removed every doubt of their situation being then on the eastern side of that pass, by which the Russians maintain a communication between their settlements in these two extensive inlets. Mr. Whidbey, however, for his further satisfaction, was very desirous of finding the road or path by which the intercourse was carried on; and although he was unsuccessful in ascertaining this, yet it did not appear to him that any particular track was necessary, as

[1] Tarr, R. S. and Martin, Lawrence, Annals Assoc. Amer. Geographers, Vol. II, 1912, pp. 38–39 and Fig. 1 on p. 27.

[2] Davidson, George, Op. cit., p. 24.

[3] Pl. II in Bull. 277, U. S. Geol. Survey, 1906.

[4] Pl. I in Bull. 327, U. S. Geol. Survey, 1907; see also Coast and Geod. Survey, Chart 8550, 1909.

[5] Cabin Creek, the largest stream from Billings Glacier.

the valley has a tolerably even surface, and was nearly destitute of any vegetable productions, and was equally passable in all directions. Its situation and character corresponded also with the description of it given by the Russians, and Mr. Whidbey's mensuration agreed nearly with the distance across as stated by them, namely, about sixteen versts."

Before 1880 when Petroff made journeys in this part of Alaska while taking the census, and 1887 when Applegate visited the head of Passage Canal, the Portage Glacier had apparently advanced and occupied Whidbey's "valley of some breadth," which contained "elevated land," and was "very free from snow and appeared to be perfectly easy of access," leaving a "tolerably even surface," "nearly destitute of any vegetable productions," and "equally passable in all directions." Petroff's description indicates that a great glacial advance had taken place, for he states that a "glacial formation forms the portage route between Chugatch bay[1] and Cook's Inlet"; and he shows the glacier on his map of 1880, much as at present. Applegate's map, based upon a close observation from his schooner in June, 1887, has the Portage Glacier joining the one from the north and ending on the land just west of the head of Passage Canal.

Although admitting a reasonable doubt that Whidbey may not have known a valley glacier when he saw one, we feel rather certain that a great advance had taken place. It is clear that the portage route used by the natives and the Russians was not the non-glacial route traversed by army parties in 1898 and 1899, although this starts from near the Russian house at the mouth of Cabin Creek on the delta of the stream from Billings Glacier. This seems certain because (a) this route is too long, and (b) within 1½ miles of Cabin Creek it goes over a 3000-foot, snow-covered pass. Such a route could not have been traversed habitually by the natives, taking their canoes with them; and that this was the case at the time of Whidbey's exploration, is indicated by Vancouver's statement that in 1794 Lieutenant Johnstone encountered some strange natives in eastern Prince William Sound and "clearly understood that the strangers had come immediately from Groosgincloose, or Cook's Inlet, and *that they*[2] *with their canoes, had crossed the isthmus overland that separates this sound from Turnagain Arm.*"

Neither does it seem probable that the natives carried canoes over the Portage Glacier as at present, for it ascends so steeply east of the divide that in 1898 the prospectors had to use ropes and pulleys in drawing their sleds up over the last ascent. Moreover, Alaskan natives are notoriously timid about trusting themselves on glaciers. None of Vancouver's maps show glaciers, even where they existed, but they are usually mentioned, so that we feel quite confident that in 1794 there was no Portage Glacier of anything like the present dimensions. Judging by his other descriptions, Whidbey would not have described, in the language quoted, a valley that contained a glacier with an ascent in the first mile to over 1000 feet, and with the terminus only three-quarters of a mile from the coast where he stood.

It therefore appears probable that before 1794 the through valley in which Portage Glacier now lies was free from glacier ice, though the tributary glaciers doubtless existed as separate ice tongues. If this interpretation of the discrepancy between present conditions and earlier descriptions is correct, there has occurred here between 1794

[1] An old name for Prince William Sound.

[2] The italics are Vancouver's.

and 1880 an advance of from one to three miles, by the forward movement and union of the two side glaciers, thus filling the pass with a double-ended through glacier.

Stagnation in Recent Years. We have no evidence of changes in Portage Glacier between 1880 and 1887, nor between 1887 and 1898. From 1898 to 1909, when the eastern end of the glacier was mapped and briefly described by Grant and Higgins, there was little change (Fig. 51), and between 1909 and 1910 there was no noticeable change. In the latter year, the glacier surface had little morainic débris and there were no conspicuous moraines. As seems to have been the case throughout its recent history, the ice at the time of our visit was only moderately crevassed. We observed no barren zone around the glacier borders, and mature trees on parts of the outwash

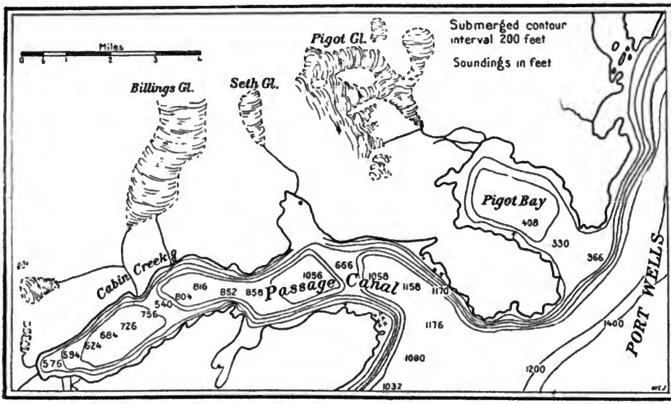


FIG. 54. MAP SHOWING INNER PORTION OF PASSAGE CANAL (PORTAGE BAY) WITH SOUNDINGS MADE IN 1910. Soundings in feet. Submerged contour interval, 200 feet. (Outline of coast and glaciers after Grant and Higgins.)

gravel plain, between the glacier and the head of Passage Canal, show that it could not have reached tidewater at any time in its recent history.

Learnard Glacier. Learnard Glacier,[1] a smaller ice tongue north of Portage Glacier, terminates within a quarter of a mile of the fiord (Fig. 54), to which it sends a stream over the outwash gravel plain in front of Portage Glacier. The glacier slopes with a very low grade and ends in the mouth of a deep, narrow valley. It is considerably encumbered with morainic débris, so that its terminus is dark.

Detached Ice Mass. Between the terminus of the glacier and the head of the fiord is a large, irregular-shaped hill, which from a distance looks like a mine dump, but which is really a detached mass of ice covered over with angular ablation moraine, including much black slate. This débris has retarded melting during the period since

[1] Named in 1910 for Lieut. H. G. Learnard.

the ice mass was separated from the main glacier. The size of the trees between Learnard and Portage Glaciers preclude the possibility that this detached ice remnant was supplied by Portage Glacier.

The Learnard Glacier has continued to melt back for the last twelve years, for while in 1909 the terminus was dark with débris, in 1898 it was white ice. The detached ice mass, however, has changed very little since 1898, for Mendenhall and Glenn described it vividly, in language which in all essential respects depicts the 1910 conditions. Yet that it is slowly wasting away is proved by its irregular surface, due to slumping and by the almost complete absence of vegetation on its surface. To show the conditions in 1898, we give Mendenhall's description, then that of Glenn, which furnishes a few additional items.

"This little glacier illustrates very well the rapidity of the ice retreat and shows us the processes whose results only remain in portions of the New England landscape. One quarter of a mile out from its present terminus is a hillock 220 feet high and half a mile long, with its longer axis parallel with the glacier front. It is now separated from this latter by an open valley paved with bowlders. At first sight this elevation was supposed to be a simple terminal moraine, but upon examination it proved to consist mostly of ice deeply covered with angular débris, which is also disseminated through it. This remnant of the glacier seems to stand near a position which the ice front occupied long enough to become covered by a sufficiently thick mantle of protective débris, so that melting was not so rapid as in the less well-protected part of the glacier just back of the front. The separation from the glacier was probably facilitated by the exit of the subglacial stream through a tunnel back of the protecting mantle. The combined melting from above and below soon removed this neck, leaving the former front isolated as it stands today. Since its isolation it has been shrinking each summer, and now occupies less than half of its original area. Around its seaward side is a belt of rough ground of slight relief covered with angular and unsorted material, which has been let down into position by the melting of the ice front. The outer rim of this zone is somewhat higher than the inner portion, giving it the form of a shallow amphitheater facing the remnant of the glacier which remains. The stability of the position of maximum advance for a short time due to the balance between flow and melting at the front accounts for the slightly greater accumulation there and the building of the low rubble wall.

These recent glacial details of topography are the more striking since they are built upon a smooth water-laid deposit of relatively fine material. This delta is of the type which usually forms before glaciers in these fiords, and gives about the only level areas to be found near sea level in a region of sharp topographic forms. At its outer margin a short distance seaward from low-tide level, the delta slopes abruptly to the profound depths so often found in these inlets."

Glenn states that in 1898 the stagnant ice block consisted of a "detached or isolated hill not far from tide-water, and contains not to exceed 2 acres of surface on its top. It is surrounded at a distance of from 30 to 50 yards from its base with a pile of rocks or bowlders, which evidently marks its original size. From this circle this moraine has gradually receded for a number of years, due to the action of the elements. Within this circle of bowlders others have been deposited, but in no regular order. The top and sides of this moraine are covered with a collection of dirt and stones to a depth

of several feet. In the northeast corner there is a circular opening that shows a clear, blue ice. At the bottom of this we discovered two caves running for an unknown distance into the ice. In front of this opening snow has drifted to a depth of 30 to 50 feet."

The preservation of such a detached ice mass for a period of over twelve years, as a result of the protection afforded by a thin cover of ablation moraine, is a matter of considerable interest.

A Cascading Glacier. The ice tongue east of Portage Glacier is a cascading glacier, heading in snowfields from which also the southern tributaries of Portage Glacier descend northward, and the Carroll, Concordia, and Milton Glaciers of Blackstone Bay, eastward. It descends part way down the fiord wall and terminates several hundred feet above sea level three fourths of a mile from the shores of Passage Canal. Mendenhall described it as a glacier which "spills over the mountain rim 2000 feet above tide in an ice cataract," and in 1910 it was much the same. A number of small streams with waterfalls drain this cascading glacier and are still building extensive outwash gravel deposits, which end in a projecting delta on the south side of the fiord, near its head. Around the borders of this cascading glacier, which is clean and severely crevassed, is a moderate-sized barren zone, showing that the glacier has recently been more extensive than at the present time. There are several small ice masses ending high on the mountain slope east of this cascading glacier.

Glacial Erosion in Passage Canal—Fiord Walls. That Passage Canal has been intensely glaciated is proved by the pronounced glacial smoothing and striation. The northern wall, near the head, has broad, eastward-descending rock benches (Pl. CXLIII), 2000 feet or more above sea level, showing a minimum height to which it has been glaciated, and proving that the ice moved eastward in Passage Canal. The northern side of the fiord, from Portage Glacier to Blackstone Bay, is much steeper than the southern, which has projecting, untruncated spurs and some islands, suggesting that the main ice stream moved with more power on the northern side. There is no essential difference in the degree of oversteepening of the northern fiord wall to the east and west of Seth Glacier cove, although the walls in the portion to the east are granite, and in the portion to the west graywacke and slate.

Cirques and Hanging Valleys. The cirque east of Seth Glacier is one of several such forms in Passage Canal, excavated by glacial erosion. The cascading glacier east of Portage Glacier, and several others near it, are in hanging valleys. The valley of Billings Glacier is apparently cut down to sea level but may have a discordant relationship to the fiord bottom. Seth Glacier is in a hanging valley, in the lip of which the stream has cut a deep rock-walled gorge.

Through Valley. The most conspicuous work of glacial erosion above sea level is in the through valley occupied by Portage Glacier (Pl. CXLIV). This has been cut well down toward sea level, forming the only low pass across the Kenai Mountains, whose other passes are 3000 feet or more high, and whose snowfields and peaks rise to a general elevation of five or six thousand feet.

Submarine Topography of the Fiord. The fiord is 576 feet deep near its head at a point within half a mile of the outwash gravel plain of Portage Glacier (Fig. 54), and the bottom slopes at an average rate of 60 feet to the mile. It is 1176 feet deep near the entrance of Blackstone Bay, ten miles to the east. Eight miles farther east, outside the entrance of Port Wells (Fig. 49), the depth is 1428 feet, the bottom slope having

decreased to 30 feet to the mile. The average bottom slope for 18 miles is, therefore, concave upward and, like a stream course, steepest at the head. The lines of soundings across the fiord opposite the mouth of Blackstone Bay, and east of Port Wells, show that it has a broad flat bottom; and the typical steep walls of the visible portion of the fiord are continued below sea level, a depth of 1170 feet being found a quarter mile from the northern shore just west of Pt. Pigot.

There are two conspicuous interruptions of the slope of the fiord bottom (Fig. 54), one near the delta of Billings Glacier, where the water is only 540 feet deep, thus rising 216 to 264 feet above the fiord bottom on either side. The other is opposite the mouth of the cove of Seth Glacier, where the water is only 666 feet deep, thus rising 392 to 390 feet above the adjacent bottom. These may be either submerged moraines or ledges between rock basins in the fiord bottom, due to glacial erosion, and information is not available for determining between these two possibilities. It is noteworthy that, as in other fiords of Prince William Sound, each of these irregularities comes near where the trunk glacier formerly received a tributary.

In addition to the visible hanging valleys along the fiord walls above sea level, there are probably also submerged hanging valleys. The cove of Seth Glacier on the northern side of the fiord, the cove southwest of it, on the opposite side of the fiord, and a number of smaller indentations are suspected to have this relationship, though soundings were not made. Blackstone Bay may have a slightly discordant relationship to Passage Canal (96 feet), though this is thought doubtful, because the combined Tebenkof and Blackstone Bay Glaciers must have nearly or quite equalled the former Passage Canal Glacier both in volume and in erosive power.

Sparseness of Glacial Deposits. There is conspicuous absence of glacial deposits in Passage Canal. The outwash gravels of the Billings Glacier valley train, with the large delta at its terminus, and the outwash flat at the head of the fiord, near Portage Glacier, are the only deposits of any size above sea level, although there is the usual irregularly distributed veneer of ground moraine. Mention has already been made of the 540 and 666-foot submarine ridges, which may represent submerged moraines due to halts of the southwestward retreating ice tongue of Passage Canal if they are not rock swells associated with eroded basins.

Vegetation. Mature forest at sea level extends up to the very head of this fiord, and there is no suggestion of recent glacial expansion excepting the small barren zone of the cascading glacier and the detached, moraine-covered ice block in front of Learnard Glacier. These are but minor exceptions, and the forest trees, which attain a diameter of a foot or two, show that the general glaciation was centuries ago.

Cochrane Bay

This fiord is just east of Blackstone Bay, and might be considered a southward extension of Port Wells, across Passage Canal. It is 1½ to 4 miles wide, 12 miles long, and its head is within 4 or 5 miles of the Applegate Arm of Port Nellie Juan fiord to the south. Glacial erosion has doubtless given Cochrane Bay much of its present form and depth, the water being 1428 feet deep near its mouth which joins Passage Canal with accordant grade, as does Port Wells to the north.

Whidbey in 1794 sketched the outlines of this bay; and Vancouver speaks of it as

"a bay about a league and a half wide, and about three miles deep in a south direction, where it was terminated by a similar boundary of ice and frozen snow as before described reaching from a compact body of lofty frozen mountains to the water's edge." Grant and Higgins infer that this description refers to either Tebenkof or Blackstone Glacier; but neither the text nor the map seem to us to harmonize with this inference.

None of the later explorers described a tidal glacier in this bay and it has no tidal glaciers at present and is free from floating ice. Three glacial streams flow into Cochrane Bay from the western side: (1) a stream from the eastern lobe of Tebenkof Glacier; (2) one from a small glacier in the mountains between the Tebenkof valley and Cochrane Bay; and (3) one from the larger Rainy Glacier (Fig. 55), which extends northeastward from the snowfields near the head of Tebenkof Glacier.

Near the mouth, at least, Cochrane Bay has thick mature forest, but since we did not see the conditions near the head, we do not know whether there is such young vegetation as to support the theory of a tidal glacier having retreated southward more than eight miles since 1794. The divide between Cochrane Bay and Port Nellie Juan is said by Grant to be less than 200 feet high, making it clear that when the Port Wells and Passage Canal ice tongues filled Cochrane Bay with ice the expanded glacier spilled over southward into the Port Nellie Juan fiord.

Culross Passage

Culross Passage, a narrow, crooked channel east of Cochrane Bay, extends southward from Passage Canal to Port Nellie Juan. It is nearly 11 miles long and from a few hundred feet to a mile wide. There is a branch inlet 2½ miles long on the western side. Culross Island, forming the eastern side of Culross Passage, rises to a height of 1 or 2 thousand feet, and the mainland on the west is even higher, but the passage is not fiord-like, having gradually-sloping walls in most places, and in some portions having very low land near its shores.

At one point, however, there are moderately high, precipitous cliffs. About half way through the passage, the soundings show a submerged divide, with depths of 66 to 78 feet, contrasting with depths of 342 and 336 feet at the northern and southern ends, the former hanging 1062 feet above the bottom of Passage Canal. This is interpreted as indicating that most of the Port Wells and Passage Canal ice moved eastward into Prince William Sound, while little ice streamed southward through Culross Passage, which was, therefore, not deepened much and has many unconsumed rock reefs. At the time of maximum glaciation, however, when Passage Canal and Port Nellie Juan were filled with great glaciers, this passage was undoubtedly full of moving ice. As seen from either Passage Canal, Culross Passage, or Port Nellie Juan, Culross Island looks as if it had been completely overridden and rounded at the expanded stage of glaciation.

CHAPTER XVIII

OTHER GLACIERS OF PRINCE WILLIAM SOUND

Geography

The remaining glaciers of Prince William Sound are those in (1) Port Nellie Juan, (2) Icy Bay, (3) the islands of southwestern and southern Prince William Sound, and (4) the fiords on its eastern side. Port Nellie Juan and Icy Bay are fiords south of Passage Canal in which terminate good-sized glaciers from the snowfields of the Kenai Peninsula. There are small local glaciers on Knight and Montague Islands, which are respectively east and southeast of Icy Bay. The fiords on the eastern side of Prince William Sound contain small-sized glaciers, which extend westward from the snowfields of the Chugach Mountains.

Acknowledgments

The National Geographic Society's 1910 expedition did much less work on the glaciers described in this chapter than in the other parts of Prince William Sound, and the discussion will, therefore, be very brief, especially as the ice tongues are much smaller than those in the fiords thus far described.

Acknowledgment has already been made of the full use of the photographs, maps, and brief reviews of results of work by Grant and Higgins, which we had for use in the field. While writing the final draft of this chapter we also had before us their preliminary account, published in 1911, and have made use of their results for the general description of the glaciers which is necessary for an understanding of the brief observations and interpretations we were able to make in 1910. The failure of our supply of gasoline made it necessary to confine our detailed field work to the glaciers in the outer portions of Port Nellie Juan and Icy Bay, and to abandon all sounding after leaving Passage Canal and Culross Passage. Our description of the ice tongues in Port Nellie Juan and Icy Bay is supplemented by the work of Grant and Higgins, and by earlier observations of Portlock, Vancouver, Seton Karr, Applegate, Glenn, Perkins, and others.

The descriptions of the glaciers of Knight and Montague Islands and of the fiords of eastern Prince William Sound, which no one has thus far visited, except for a hurried trip up Port Fidalgo by Schrader, are based upon our views from a distance.

The charts of the United States Coast and Geodetic Survey furnish data for the interpretation of submarine topography, the latter topic being partly postponed for discussion in the chapter on the glaciation of Prince William Sound.

Port Nellie Juan

Topography. Port Nellie Juan, which is connected with Passage Canal by Culross Passage, consists of three parts:—(a) an outer portion, trending northeast-southwest

from Prince William Sound, with a length of 12 miles and a width of 2 to 4 miles; (b) a middle portion trending nearly at right angles with the last and having a length of 6 miles and a width of about 2 miles; and (c) an inner portion, called Applegate Arm, extending at right angles to the last and nearly parallel to the outer division, with a length of 13½ miles, and a width of 2 or 3 miles.

At the southern end of the outer portion is the cove containing Nellie Juan Glacier, and Blue Fiord, at the head of which is Ultramarine Glacier. A fiord called McClure Bay lies further northeast. With the exception of Contact Glacier on the southern side there are no ice tongues close to the shore of the second section of the fiord, which has several indentations. Applegate Arm has three ice tongues on the western side and five on the eastern. The Kenai Mountains near this fiord rise to heights of four or five thousand feet, with peaks that are even higher.

Observations of Glaciers. The observations of the ice tongues of Port Nellie Juan began in May, 1887, when Applegate discovered and explored this fiord, making a map which was first reproduced by Davidson.[1] Applegate published no description of the glaciers.

In August, 1908, Grant and Higgins spent two days in this inlet, making a map (Fig. 55), taking many photographs, and making the observations which are quoted in this chapter.[2]

The junior author spent August 6 to 8, 1910, in Port Nellie Juan, visiting the Nellie Juan Glacier and seeing Ultramarine and Cotterell Glaciers from a distance. As already stated, we were unable to make soundings or to visit Applegate Arm.

Nellie Juan Glacier—The Ice Tongue. Nellie Juan Glacier rises in unexplored snowfields of the Kenai Mountains (Fig. 55), flowing northeastward to a cove on the southern side of the fiord, with a known length of over 3 miles and a width at the terminus of a mile. West of it a smaller glacier descends from the same snowfield. Nellie Juan Glacier is only moderately crevassed, has a low terminal slope, and two miles from the terminus is fed by two good-sized tributaries, the larger branch being on the northern side. The cove in which this glacier terminates is about 2½ miles long, and 1½ to 2 miles wide. Although the glacier extends down to this cove, the terminal ice cliff of 1910 did not reach the sea at low tide except in a small V-shaped indentation near the eastern margin, the only point where icebergs were then discharged. At high tide the water bathed most of the terminus of the glacier, which was precipitous though not very high; but at low tide the glacier front was separated from the cove by a narrow strip of sand, across which several streams flowed, the chief one being near the western edge, where the glacier terminus rested on a granite hill.

Retreat of Glacier. When first mapped by Applegate in 1887, this glacier is represented as terminating in the sea, its ice cliff facing the northeast. There was slight retreat from 1887 to 1908, when Grant and Higgins showed the terminus on the land, much as in 1910. Their description states that the end of the glacier "rests on a gravel beach, most of which is covered by high tide; and near the center of the front the ice is bathed

[1] Applegate, S., map reproduced in Davidson's Glaciers of Alaska that are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc., Geog. Soc. of the Pacific, Vol. 3. 1904, pp. 26–7 and Map XI.

[2] Grant, U. S. and Higgins, D. F., Glaciers of the West Coast of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLIII, 1911, pp. 409–413; preliminary notes in H. F. Reid's Variations of Glaciers, Journ. Geol., Vol. XVII, 1909, p. 671; maps in Bull. 284, U. S. Geol. Survey, 1906, Fig. 4, p. 79; in Bull. 379, *Ibid.*, 1909, Pl. IV, facing p. 88, and in Bull. 443, *Ibid.*, 1910, Pls. I and II.

by low tide water. On both sides of the lower part of the glacier is a distinct bare zone of smoothed granite, and this bare zone, which is 100 to 500 feet in width, ends abruptly at the edge of a forest covered tract. This zone is prominently developed on a granite knob, almost an island, at the west side of the glacier front. Crossing the top of this knob is a small moraine . . . from 1 to 10 feet in height and 5 to 30 feet in width. This moraine contains decaying fragments of trees and just to the north of it is an area of scattering trees, some of which are a foot in diameter. To the south of the moraine is some vegetation,—moss, grass, alders 5 feet high, and a few spruce trees 4 feet high. Most of the vegetation disappears half way from the moraine to the ice front. From the extreme summit of the above granite knob the nearest point of the moraine is 48 feet distant in a direction S. 10° W. From the same summit the extreme front of the glacier is 500 feet distant in a direction S. 13° W.

"The moraine noted above marks the farthest advance of the ice since the growth of the present forest, i. e., for a century, and most probably for a few centuries. The date of this maximum historical advance, is at a minimum, twenty years, and probably the actual date is considerably longer ago than twenty years."

Two years later our comparison of photographs showed little retreat since 1908. The glacier surface was quite smooth and apparently still wasting, and a smaller portion of the ice front was bathed by low tide. The marginal stream on the eastern side was of great size and was building a steeply-sloping delta whose area increased greatly from 1908 to 1910. Grant's conclusion that the advance in association with the terminal moraine had taken place at least twenty years before 1908 would seem to associate the advance with the tidal condition of the glacier when mapped by Applegate in 1887. In connection with the long stand of the glacier terminus, with only about 500 feet of retreat in 23 years, there has accumulated an extensive terminal deposit of gravel, sand, and clay, laid down in large part below sea level. The weak terminal moraine on the land, the broader terminal deposit in the sea, and the barren zone are all conspicuous phenomena in connection with the history of Nellie Juan Glacier from 1887 to 1910.

The Outwash Fan. West of the rock hill a narrow lobe of the glacier sends a stream northward across an extensive outwash gravel fan, the margin of which is submerged for ¼ mile at low tide. This fan, which is over ½ mile wide, has tied Nichols Island to the mainland, but on the northern edge, near the island, there is a channel connecting the main cove with the narrow inlet to the west of Nichols Island. A slough of clear water flows across this flat even at low tide, and enters the bay in front of Nellie Juan Glacier. Its current is distinguishable for several hundred feet offshore in the midst of the milky-white glacial waters of the rest of the cove. Another of the streams from the western margin of the glacier flowed across the fan to the branch of the fiord west of Nichols Island. This portion of the fan completely surrounds a rock islet. The surface of the fan is barren, evidently because the glacial streams have occupied its surface continuously. It has an unusually steep slope and parts of it are made up of very coarse bowlders. In recent years, with the retreat of the glacier and diminution of débris supply, some of the streams have incised moderately-deep gulleys in the fan.

Ultramarine Glacier. This glacier terminates in Blue Fiord, which is 4½ miles long, ½ to 1 mile wide, and very precipitous. The length of the Ultramarine Glacier is unknown, only a mile of it having been mapped, and at the terminus it has a width of about ¾ of a mile (Fig. 55). Like Nellie Juan Glacier, it is fairly clean and moderately

crevassed. It is not tidal, but is separated from the fiord by a narrow strip of outwash gravels.

Applegate, the only one who has been close to the glacier, so far as we know, indicates on his map that in 1887, when he went within less than half a mile of it, there were visible reefs in front of the terminus, which was then tidal.

Grant and Higgins, after their visit in 1908, stated that "the glacier comes within about a quarter of a mile of tide water and the western part of the front extends farther forward than the eastern two-thirds and rests on a glacial flat. The eastern part of the front

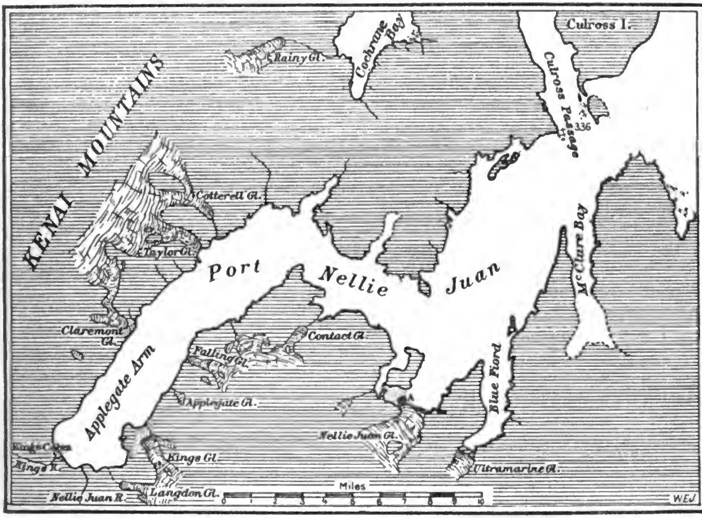


FIG. 55. PORT NELLIE JUAN AND ITS GLACIERS IN 1908 (AFTER APPLEGATE, AND GRANT AND HIGGINS).

rests on a rock ridge about 300 feet above the sea. On this ridge there is a marked bare zone, and also one on the side of the glacier. The front of the glacier was not visited, but at a distance this bare zone appears as if the ice had retreated from it in the last two or three years. Applegate's map indicates that the glacier in 1887 reached to tidewater along its whole front. The forest in front of the eastern part of the glacier shows that this could not have been the case, although the western part may have reached tide water at that time, but even this is doubtful. Our observations on this glacier were made at a distance of about a mile and a half."

Since we saw Ultramarine Glacier in 1910 from a distance of several miles we cannot say whether it had changed any in the last two years. It was evident, however, that there had been no great change in the way of advance or retreat.

Glaciers of Applegate Arm. The glaciers of Applegate Arm are Cotterell, Taylor, and Claremont Glaciers on the western side, and Langdon, Kings, Applegate, Falling, and smaller ice tongues on the eastern side. The ice tongues of the western side of Applegate Arm are fed from the snowfields which also supply ice to Tebenkof Glacier and the ice tongues at the head of Blackstone Bay. Several of these glaciers were mapped by Applegate, who went close to them when he discovered this fiord, in 1887. They were surveyed more in detail by Grant and Higgins in 1908 (Fig. 55), but up to the present time, with the exception of a brief note, quoted below, they have published no description of the outwash gravel fans and barren zones, which they mapped, together with a terminal moraine of Claremont Glacier. They stated that when they visited this fiord they did not examine in detail the glaciers on its western side.

"South of the central part and east of the southern part of Port Nellie Juan is a snowfield of unknown but considerable extent. Several glaciers flow north and west from this field, and . . . Falling Glacier reaches tide water. On the west side of the southern part of the port (Applegate Arm) are other glaciers, one of which, the Taylor, reaches tide water. The head (of Applegate Arm) is shallow and the waters very muddy. This is due to streams from the adjoining glaciers, and especially to . . . Kings River, entering the head of the port . . . charged with glacial silt (coming) from one of the largest ice covered areas of the Kenai Peninsula."

The National Geographic Society's party saw only Cotterell Glacier in 1910 and this at a considerable distance, so we have no information as to the behavior of the glaciers of Applegate Arm from 1908 to 1910. The largest glacier of Applegate Arm, which supplies the water of Kings River, has not yet been mapped or studied.

Glaciation of Port Nellie Juan. In this broad, deep fiord the evidence of the former extension of the glaciers is clear, the glaciated fiord walls, especially the precipitous northwestern side of Blue Fiord, the plucking of the well-jointed granite of the rock hill at the terminus of Nellie Juan Glacier and on Nichols Island, and the glacial striae, making it plain that the whole fiord was formerly filled by an ice tongue which flowed into Prince William Sound.

Glacial deposits are of small extent, and the fairly thick, mature growth of vegetation (for example, on Nichols Island and on the mainland near Nellie Juan Glacier) makes it clear that the last greatly expanded stage of these glaciers was more than a century ago.

The small barren zones near Nellie Juan and Ultramarine Glaciers, and near the ice tongues of Applegate Arm, mapped and described by Grant and Higgins, are interpreted as evidence of a slight advance, two decades or more ago, similar to the advances in the fiords to the north.

The extent to which glacial erosion has deepened and otherwise modified the fiord is likewise unknown. Judging by the absence of many reefs and of kelp away from the shores, it is assumed that here, as elsewhere, the fiord has been deepened throughout by glacial erosion.

Icy Bay

General Description. The entrance to Icy Bay, which is connected with Prince William Sound both to the north and south by a broader fiord called Knight Island Passage, is between Chenega Island and Point Countess. The outer part of Icy Bay (Pl. XCIII) trends northwest-southeast for 8 miles, with a width of 3 to 4 miles; while the inner

part extends at right angles to it with a length of about 13 miles and a width of from ½ to 1½ miles. Facing the entrance is Chenega Island, north of which Dangerous Passage also connects Icy Bay with the northern portion of Knight Island Passage. The inner part of Icy Bay has a cove on the southern side which is separated from Port Bainbridge by a narrow isthmus from one half to three fourths of a mile in width (Fig. 56). The head of Icy Bay, to the southwest of this cove is less than half a mile wide and is terminated by a tidal ice tongue called Tiger Glacier. On the northern side of inner Icy Bay is a larger indentation nearly two miles in length and from one half to two miles wide, and partly cut off from Icy Bay by a high, rocky peninsula. Chenega, Princeton, and Tiger's Tail Glaciers terminate in this indentation, which is called Nassau Fiord. Small icebergs from this bay drift out into Knight Island Passage, sometimes being seen as far east as Latouche Island, along a channel regularly traversed by steamers. The depths of water in Icy Bay are unknown except at the mouth, in Knight Island Passage, where the fiord is 1000 to 1350 feet deep,[1] and in Icy Bay opposite the mouth of Nassau Fiord where it is 300 feet.[2] The fiord walls are steep and the adjacent mountains rise to heights of several thousand feet above sea level.

Observations of Glaciers. The information about the glaciers of Icy Bay is based upon observations by Portlock [3] in 1787, Vancouver [4] in 1794, Seton Karr [5] in 1886, Applegate [6] in 1887, Glenn [7] in 1898, Grant and Higgins [8] in 1908, Perkins [9] in 1909, and the National Geographic Society's expedition [10] in 1910.

Glaciers of Nassau Fiord—Chenega Glacier. Chenega Glacier is known to be at least 3 miles long, plus a continuation in unexplored snowfields, and is fed by two large tributaries, and at least one small one. It terminates in Nassau Fiord with a vertical ice cliff nearly half a mile in width and 135 feet high. It is a clean white ice tongue with no medial moraines and descends a steep slope of 500 or 600 feet in the last quarter of a mile, with severe crevassing. Above this the glacier has a more moderate slope. Many icebergs are discharged from this, the most active tidal glacier in Icy Bay. Around its terminus, and on a nunatak between its two chief tributaries, is an extensive barren zone; and the absence of vegetation throughout most of the lower slopes of Nassau Fiord indicate recent extensive retreat.

Historical evidence, reviewed later, indicates that Chenega Glacier, joining the ad-

[1] U. S. Coast and Geod. Survey, Chart No. 8550.

[2] Soundings by Capt. W. P. S. Porter of the steamship *Yucatan* which took the Perkins party into Icy Bay in 1909. Personal communication from Mr. George W. Perkins.

[3] Portlock, Nathaniel, Voyage Round the World, London, 1789, p. 240.

[4] Vancouver, George, Voyage of Discovery to the Pacific Ocean and Round the World, Vol. V, London, 1801, pp. 304–305; maps republished by Davidson (see below) Maps IV and V.

[5] Seton Karr, H. W., Shores and Alps of Alaska, London, 1887, pp. 228–229.

[6] Applegate, S., quoted in Davidson's Glaciers of Alaska That are Shown on Russian Charts or Mentioned in Older Narratives, Trans. and Proc. Geog. Soc., Pacific, Vol. 3, 1904, p. 23.

[7] Glenn, E. F., War Dept., Adj.-Gen. Office, No. XXV, 1899, map in pocket.

[8] Grant, U. S. and Higgins, D. F., Glaciers of the West Coast of Prince William Sound, Bull. Amer. Geog. Soc., Vol. XLIII, 1911, pp. 410, 414–416; Tidewater Glaciers of Prince William Sound and Kenai Peninsula, Bull., U. S. Geol. Survey (in preparation); also in H. F. Reid's Variations of Glaciers, Journ. Geol., Vol. XVII, 1909, pp. 670–671; also Maps in Bull. 284, U. S. Geol. Survey, 1906, Fig. 4 on p. 79; *Ibid.*, Bull. 379, 1909, Pl. IV facing p. 88; *Ibid.*, Bull. 443, 1910, Pl. I facing p. 10 and Pl. II, in pocket.

[9] Perkins, George W., Personal communication.

[10] Martin, Lawrence, Nat. Geog. Mag., Vol. XXII, 1911, pp. 551, 555; Journ. Geol., Vol. XIX, 1911, p. 458.

jacent ice tongues, was much more extensive and retained essentially the same position between 1787 and 1898, and that it retreated a long distance between 1898 and 1908. In the latter year the glacier was studied and mapped by Grant and Higgins,[1] when they spent one day in Icy Bay, making the map reproduced as Fig. 56, taking photographs and discovering the retreat of two to three miles. They quote a native tradition that this glacier extended to the mouth of Icy Bay about a century before and interpret this as meaning the mouth of Nassau Fiord because of the relationships of forest and barren zones.

In 1909 the G.W. Perkins party visited Icy Bay and proposed that this ice tongue be called Princeton Glacier; but, since Grant had previously called it Chenega, the name Princeton is applied by Grant to the glacier east of it. Perkins also named the Tiger's Tail Glacier, the Tiger Glacier, and Nassau Fiord; and one of the officers of his ship determined the height of the ice cliff of Chenega Glacier.

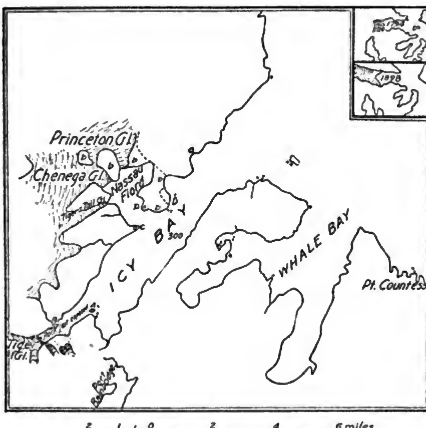


FIG. 56. THE GLACIERS OF ICY BAY IN 1794, 1898, AND 1908.
Barren zone shown by dashed line and letter b. Soundings in feet. (After Vancouver, Glenn, and Grant and Higgins.)

The National Geographic Society's 1910 party spent portions of two days at Chenega Glacier, and found the ice front in about the same position as in 1908. A comparison of photographs, taken from one of the Grant and Higgins' photographic stations (c, Fig. 56), showed in the two years no significant change, although conditions along the eastern edge, and an especially marked increase in the height of the terminal cliff, suggested the possibility of a slight advance being then in progress (Pl. CXLV, A).

Princeton Glacier. Princeton Glacier, of about the same size as Chenega Glacier and fed from adjoining snowfields, is no longer tidal. The terminus of the eastern margin is so encumbered with ablation moraine, that this side of the glacier appears dirty. About a half mile back from the water's edge, there is a medial moraine near the eastern margin which, however, extends diagonally across the glacier, so that at the terminus, it is close to the western margin.

The lower end of the Princeton Glacier slopes gradually, in contrast with the frontal precipice and steep slope of Chenega Glacier, and the surface is very little crevassed.

[1] Bull. Amer. Geog. Soc., Vol. XLIII, 1911, pp. 410, 414–416.

As far back as can be seen from Icy Bay the glacier has a moderate slope, suggesting that it comes from a considerable distance. It is possible that its snowfields join those of Nellie Juan and Ultramarine Glaciers, 10 or 12 miles to the north.

There are barren areas around the terminus of Princeton Glacier, which seems also to have had a more expanded stage over a century ago, and to have then coalesced with Chenega Glacier to form the expanded Icy Bay Glacier.

When mapped by Grant and Higgins in 1908 three quarters of the glacier terminus was separated from the water's edge by a low line of morainic or rock hills, but the western margin may have been tidal. Between August 5, 1908, and August 9, 1910, the glacier was wholly inactive, continuing to waste so rapidly that the areas covered by ablation moraine increased decidedly in amount, the ablation moraine of the eastern side practically coalescing with the diagonal medial moraine. By that time, also, there was sufficient retreat of the margin so that it nowhere reached the sea, even the western edge being separated by narrow gravel and sand strips. It is possible that at high tide the water covered this fringe and touched the ice edge in the western part; but there was no terminal cliff, and no icebergs were discharged.

Tiger's Tail Glacier and Smaller Ice Tongues. Tiger's Tail Glacier is a slender ice tongue, extending down over the lip of a hanging valley as a cascading glacier, and reaching the water's edge, so that a few icebergs are probably discharged. It was first mapped by Grant in 1908.

The small ice tongues to the southwest of Tiger's Tail Glacier are fed from adjacent névé fields. One glacier conspicuous from Nassau Fiord terminates in a hanging valley 800 to 900 feet above the sea, to which it sends two streams which flow down either side of a broad mountain spur. The terminus of this glacier is more or less veneered with moraine and in front of it, on the lip of the hanging valley, are good-sized spruces and hemlocks and a thick growth of willow and alder, making it extremely improbable that this glacier could have participated in the expanded condition of 1794 when Chenega, Princeton, and Tiger's Tail Glaciers filled Nassau Fiord. The other ice tongues farther to the southwest terminate high above sea level, several of them in hanging valleys, and from all of them streams with waterfalls descend the fiord wall to the northwestern side of Icy Bay.

Tiger Glacier. Tiger Glacier, at the head of Icy Bay, seen by us only from the mouth of Nassau Fiord, is shown in the map (Fig. 56), photographs, and brief description by Grant and Higgins, to be fed by two good-sized tributaries. It has been mapped for only a little over a mile, although the photographs show clearly that it heads some distance farther to the southwest. The adjacent snowfields also feed several small glaciers to the east of it, which terminate several hundred feet above the fiord and supply streams which cascade down the fiord wall. Tiger Glacier, which in the portion seen is severely crevassed, terminates with a steep slope, at the end of which is a vertical cliff less than half a mile wide, from which icebergs are discharged. Near the northern margin bare rock was exposed beneath the terminus of the glacier for over one-third of its width in 1908. A photograph by the Perkins party in 1909 shows that Tiger Glacier had advanced slightly, covering a prominent rock which was visible the year before. There was no marked continuation of this advance up to August 9, 1910.

Retreat of the Icy Bay Glaciers. The first historical suggestion of a glacier in Icy Bay results from the visit of Portlock to Prince William Sound in 1787. In July of that year

Portlock's whaleboat and yawl cruised in the vicinity of what seems to be Knight Island Passage, while looking for the Passage Canal route to Cook Inlet, which they seem to have thought to be a river. They state that "On getting over to the South West shore, they met with great quantities of drift-ice, coming as they supposed, out of that opening, and at the same time heard a constant jumbling noise resembling the breaking up of ice in a large river." The officer in command encountered fog and did not push on or see the glacier, whose presence he seems not to have suspected. The map is indefinite, but the description of icebergs and noises indicates that a tidal glacier in Icy Bay, extended as far or farther down the fiord than seven years later when Vancouver's expedition entered these waters.

The first description and map of the Icy Bay Glacier was made in 1794 by Vancouver's expedition. Lieutenant Whidbey states that in going up outer Icy Bay from Point Countess they passed "many large pieces of floating ice, which were in great abundance in this part of the sound." He describes inner Icy Bay, which they saw on June 4th, as "a bay on the western shore about a league wide, and about four and a half miles deep terminated by a compact body of ice that descended from high perpendicular cliffs to the water side, and surrounded by a country composed of stupendous lofty mountains covered with snow." The map accompanying Vancouver's account shows the terminus of the glacier at a point in Icy Bay somewhere east of the head of Port Bainbridge and probably near the entrance of Nassau Fiord. The representations of the shorelines on this map (see inset, Fig. 56) are sufficiently detailed for us to identify the location with some certainty.

On October 20, 1886, Seton Karr visited the native village at Chenega near the mouth of Icy Bay, and he has published a sketch of Icy Bay as seen from Chenega Island, showing a glacier and snowfields in the background. Since Tiger Glacier could not be seen from Chenega Island, and since Chenega-Princeton Glacier cannot be seen from that point at present and would not be visible unless they were sufficiently expanded to extend outside the mouth of Nassau Fiord, this picture is interpreted as proof that the Chenega-Princeton Glacier remained in the expanded position of 1787–1794 up to at least 1886. Seton Karr speaks of Icy Bay near Chenega village as "a broad bay covered with small icebergs" and says that "close at hand several glaciers descend into the sea from . . . low flat snowfields." He could not have made the latter statement on the basis of what one may now see at Chenega.

After his visit in May, 1887, Applegate, quoted by Davidson, referred to this ice tongue as "a fine glacier, coming down to the water." He seems to have made no map of Icy Bay so that we do not feel sure from his description alone exactly where the glacier ended at that time, though, as stated above, Seton Karr's sketch of the previous year fixes the approximate location of the terminus.

The army expedition of 1898 under Glenn published a map (see inset, Fig. 56), showing Icy Bay with a glacier which terminated east of the head of Port Bainbridge as in 1794, and with an outline different enough from that sketched by Whidbey to suggest that it was located by a new survey, as the roughly-contoured fiord walls also suggest. No Icy Bay Glacier is mentioned in Glenn's account and the map may possibly be based on some earlier survey, later than 1794. If made in 1898 it shows that the Icy Bay Glacier front remained in essentially the same place from 1794 to 1898, changing only in shape. The map shows "floating glacier ice" in Knight Island

Passage outside Icy Bay. We feel that Glenn's map, corroborated by Seton Karr's sketch, strongly suggests a maintenance of the 1794 position for over a century.

By 1908, as Grant states specifically [1] and as his map shows, there had been a great retreat; but he does not discuss the question whether it was a gradual retreat from 1794 to 1908, or a rapid one during the last few years of this period. He says "a traverse of the shoreline of this bay in 1908 shows it to be about 11 miles in length with a tide-water glacier at its head. On the north side of the bay, 6 miles from its head, is a smaller bay, nearly two miles in length; and at the head of this bay are two tide-water glaciers. The description of Whidbey, who was attached to Vancouver's exploratory expedition of 1794, states that this bay was four and a half miles deep and was terminated by a perpendicular cliff of ice. This would seem to indicate a retreat of ice in the axis of Icy Bay of some six and a half miles from 1794 to 1908."

Relationships of Forest Growth. Our study of the vegetation in Icy Bay is interesting in connection with the description by Whidbey in 1794 and the position of the glacier as indicated on his map. Northeast of Nassau Fiord the forest is everywhere thick and mature, containing trees from 24 to 32 inches in diameter. The annual rings in a number of them were counted in 1910 and three trees about 24 inches in diameter were found to be 113, 120 and 122 years old. There were trees of about the same diameters, and about as thickly set, all along tne coast up to and including the small island northeast of the entrance to Nassau Fiord. These are not stunted trees, even those nearest the glaciers being well-developed. As Whidbey's visit was 116 years before ours of 1910, the presence of trees 120 to 122 years old proves clearly that the glacier could not have extended quite as far to the northeast in 1794 as the site of these trees.

Along a sharply defined line near the entrance of Nassau Fiord, however, this mature forest ends, and the interior shores of Nassau Fiord, the mountain between Princeton and Chenega Glaciers, and the shores of Icy Bay from that point southwestward, have only scattered trees, none of which seem to be more than a score of years old. On the first prominent rock point inside of Nassau Fiord (near Photo. Sta. D, Fig. 56) the higher part of the barren zone has scattered willows, alders, and young spruces and hemlocks at an elevation of about 200 feet above sea level, the oldest one counted having 22 annual rings, though some of the others may have been even older. Nearer sea level the slopes are absolutely barren. This suggests that the ice front observed by Whidbey was maintained up to rather recent times, as we interpret Glenn's map to show also.

On and near the col between Icy Bay and Port Bainbridge there are scattered conifers which from a distance seem to be of good size. We did not ascertain whether these extend down to sea level or not, nor how old they are. If it should prove that there are mature trees at sea level in this part of Icy Bay, it would suggest either (a) a forest advancing over the low col from Port Bainbridge, before the Tiger Glacier portion of the dismembered Icy Bay Glacier had retreated to its present position, or (b) that the expanded ice tongue of 1787–1794 was supplied wholly by the Chenega and Princeton Glaciers which emerged from Nassau Fiord and filled this part of Icy Bay without being joined by Tiger Glacier, the upper part of the inlet being either an arm of the sea or a lake.

This latter interpretation, made in 1910, is at variance with the statement of Grant,[2]

[1] Grant, U. S., Journ. Geol., Vol. XVII, 1909, pp. 670–671.

[2] Op. cit., p. 671.

quoted on a previous page and in our earlier publications, that the ice in the axis of Icy Bay retreated six or seven miles from 1794 to 1908, for this involves the participation of Tiger Glacier in the expansion of 1794. As we revise this text, however, we have the later statement, published in 1911 by Grant and Higgins,[1] that in 1794 "it is very probable that the glaciers in Nassau Fiord . . . completely filled that fiord and extended out into, but not across, the main part of Icy Bay." If this means that Tiger Glacier did not participate in the expansion of the Chenega-Princeton ice tongue it is quite in accord with the distribution of vegetation.

One of the most interesting features in relation to forest growth in this region is the shape of the barren zone on the peninsula east of Nassau Fiord, between that indentation and Icy Bay. This shows clearly that while the expanded Chenega and Princeton Glaciers filled Nassau Fiord and extended over the narrow northeastern portion of the peninsula, the higher southeastern side of the peninsula and the low island there, which are thickly forested, were not covered by the glacier. The dashed line (Fig. 56), showing the boundary of the barren zone indicates, therefore, that the expanded Chenega-Princeton Glacier was tidal in Icy Bay in the cove east of this peninsula as well as opposite the mouth of Nassau Fiord; but that the two tidal termini did not coalesce and transform the higher part of this peninsula into a nunatak, nor override it completely.

Glacial Modifications in Icy Bay. Icy Bay, like the other fiords of Prince William Sound, has been intensely glaciated, giving it very steep walls and large areas of bare, striated rock often having the roches moutonnée form. Hanging valleys are also present, the one containing the small ice tongue south of Tiger's Tail Glacier forming an especially good illustration.

Icy Bay, even more than the fiords to the north, is remarkable for the absence of glacial deposits above sea level, the chief deposits observed by us being scattered, glacial bowlders on the bare rock surfaces and a thin veneer of ground moraine in places. There is a narrow strip of outwash gravel which separates Princeton Glacier from the fiord, and a very perfect, though minute, terminal moraine ridge, in the second cove east of Nassau Fiord on the northern side of Icy Bay (a, Fig. 56). It is made up of stony till and extends across the valley, at the head of the cove, as a single, narrow-crested ridge, with a height of between 20 and 30 feet. It seems to have been built by a projecting lobe from the Icy Bay Glacier rather than by a retreating glacier in this small valley.

Glaciers on the Islands of Prince William Sound

Knight Island. Knight Island, in western Prince William Sound, east of Icy Bay and Port Nellie Juan, is 26 miles long and 7 or 8 miles wide (Pl. XCIII). Its coast is deeply indented by fiords and its surface is very rugged. The peaks are 2000 to 3180 feet high, and since many of them are snow-capped, it is probable that their névé fields give rise to a number of minute glaciers. None of these have been shown on a map, and only one of them, a cascading glacier on the southern side of the island near Mummy Bay, was seen by the National Geographic Society's expedition. The large number of cirques show clearly that extensive local glaciers have existed in the past. Their relationships to the expanded ice sheet of Prince William Sound are discussed in a later chapter.

Chenega Island. Chenega Island, between Knight Island and Icy Bay, is about 8

[1] Bull. Amer. Geog. Soc., Vol. XLIII, 1911, p. 414.

miles long and 3 miles wide, and is much less rugged than Knight Island. All of the island is high, the summits rising to elevations of 1800 to 2000 feet, and retaining snow as late into the summer as August; but the island is not known to contain any glaciers, although we suspected the presence of a small glacier in one of the snow-filled cirques on the northwestern side, near Dangerous Passage. There are a few large cirques, and the rugged group of summits apparently have been sharpened by local glaciation.

Latouche Island. Latouche Island, south of Knight Island, with its southern end in the Pacific Ocean, is 12 miles long, 2 to 3½ miles wide, and has peaks 1730 to 2225 feet high, some of which retain snow rather late into the summer, though none of them are known to support glaciers at present. In former times, however, on the northwestern side of Latouche Island, there were good-sized local glaciers, which excavated great cirques from one to two miles wide, described and mapped by Grant and Higgins.[1] The junior author spent several hours on Latouche Island in 1904 and again in 1910. So much snow remained in the cirques in August and September of each year that it seems quite likely that the larger cirques still contain small, inactive, ice masses.

Montague Island. This, the largest island in Prince William Sound, has a total length of over 50 miles and a width of from 3 to 10 miles. It extends northeast and southwest across the mouth of the sound, nearly parallel to the trend of Latouche and adjacent islands, being separated from Latouche and Knight Islands by Montague Strait. The island is mountainous throughout, the highest peaks, including Montague Peak near the eastern end, lying on the ocean side. The mountains rise to elevations of from 3500 to 4500 feet, and one of the higher peaks near the middle of the island reaches 5641 feet.

The higher summits of Montague Island throughout the whole of its length are snow-capped during the entire year, and probably feed a number of moderate-sized glaciers, those on the side toward Prince William Sound being the largest. One or two of these glaciers, in the region south of Green Island, were seen by us from a distance in 1910; but since Montague Island is largely unexplored none of these glaciers have as yet been shown upon a map.

Extensive local glacial erosion on Montague Island is indicated by scores of cirques in the higher parts of the island and by hanging valleys, visible from Prince William Sound. There are also smaller cirques on the southwestern side of the island, showing that local glaciers formerly descended toward the Pacific Ocean.

Hinchinbrook and Hawkins Islands. Hinchinbrook and Hawkins Islands, northeast of Montague Island, separate the eastern part of Prince William Sound from the Pacific. Hawkins Island is 22 miles long and 3 to 4½ miles wide. Its highest peaks are 1846 to 2025 feet high and are not known to contain any glaciers, though cirques give evidence of former local glaciation.

Hinchinbrook Island is 22½ miles long and 13 miles wide. The northern part of the island has a low foreland, south of which are peaks from 1427 to 2910 feet high. Several of these peaks retain snow throughout the summer, and there are small ice masses, spoken of by Seton Karr [2] as "comparatively insignificant glaciers." None of these ice tongues has ever been mapped.

[1] Grant, U. S. and Higgins, D. F., Bull. 443, U. S. Geol. Survey, 1910, p. 19, Fig. 3, on p. 30, and Pl. XII, in pocket.

[2] Seton Karr, H. W., Shores and Alps of Alaska, London, 1887, p. 193.

There are many cirques in the high, northern part of the island, showing that there has been extensive local glaciation here. The southern part of the island is plateau-like, having no rugged topography but a broad, rolling upland with rounded summits, the highest of which are Signal Mountain, 1546 feet high, on the southwestern end near Port Etches, some elevations of 2260 feet near the middle of the island, and summits 1709 feet high near the eastern end. This southern half of the island has extensive cirques, a number of conspicuous ones cutting into the plateau-like upland, with valleys extending northward to Port Etches (Pl. CLXXXIII).

Glaciers of Eastern Prince William Sound

Among the fiords of eastern Prince William Sound, Valdez Arm and Port Valdez in the northeastern corner have already been described (Chapter XIII). The fiords south of this are Port Fidalgo, Port Gravina, Sheep Bay, Simpson Bay, Orca or Cordova Bay, and Orca Inlet.

Port Fidalgo—Topography. Port Fidalgo, which has also been called Fidalgo Bay, is the largest of the fiords on the eastern side of Prince William Sound, south of Valdez Arm. It is 25 to 38 miles long, as shown by different maps, and 5½ miles wide at the mouth decreasing to two miles or less inside the fiord. Its general course from Prince William Sound is east-northeast, but near the head it turns at a sharp angle and trends southeastward. There are a number of tributary bays, including Snug Corner Cove, Two Moon or Bowie Bay, Irish Cove, and Whalen Bay on the southern side, and Fish, Landlocked, and Boulder Bays on the north. The latter is also connected with Valdez Arm through Tatitlek Narrows, behind Bligh Island, which lies north of the entrance to Port Fidalgo. Goose Island lies south of the entrance.

This fiord extends far back into the Chugach Mountains. Near the entrance the fiord walls rise 1600 to 3000 feet close to the water's edge, back of a narrow fringe of foreland, but there are many peaks rising to greater heights, among them being Copper Mountain, 3830 feet, and Mt. Denson, 5836 feet. There has been no detailed topographic survey in the eastern half of Port Fidalgo, but the fiord walls there seem to be still steeper and there are peaks rising to 3721 feet, 5558 feet, and 5182 feet, while there are many higher ones in the Chugach Mountains still farther to the east.

Contracted Condition of Glaciers. The glaciers of this fiord are very small and long ago receded far into the mountains, from which they have not projected far enough to be tidal since before 1778. A great many mountains about the eastern end of Port Fidalgo are snow-capped throughout the summer, and there are probably many moderate-sized glaciers, besides the few small ones visible from the entrance to the fiord. The only ones thus far shown upon a map are the glacier on Mt. Denson and a little ice tongue terminating half a mile or so northeast of the head of Port Fidalgo. Doubtless the unexplored mountain valleys contain fair-sized glaciers, fed by the extensive snowfields of this portion of the Chugach Mountains.

This fiord was visited by Cook in 1778, Meares in 1786, Dixon in 1787, Fidalgo in 1790, and Johnstone in 1794, but none of these explorers mention floating ice, from which we infer that there was no tidal glacier within it in the latter part of the eighteenth century. Their descriptions of the snow-capped mountains suggest conditions not unlike those at present.

Petroff intimates [1] that some of the early Russian accounts to which he had access, suggested a tidal glacier in Port Fidalgo, and in one place he speaks of the "tremendous glaciers" in Port Fidalgo. As he seems to have made a mistake with regard to the Valdez Glacier being tidal at the same time and Columbia Bay having no tidal glacier, it is quite possible that he is in error about the fiord in which these glaciers were located.

In May, 1898, Port Fidalgo was explored by F. C. Schrader [2] who found no tidewater glaciers, but reported that "the shore is mountainous throughout, some of the mountains rising to a height of 4000 feet, into a region of perpetual snow and glaciers, especially on the north and east. Two mountain streams of considerable size, and apparently of glacial origin enter the inlet near the head—one from the north and one from the east. They have done much toward silting up the inlet, and flow over extensive deltoid deposits of gravel and mud flats at their mouths, especially at low tide." In this year Emil Mahlo made a rough contour map of Port Fidalgo, without mapping any of the glaciers.

The U. S. Coast and Geodetic Survey parties [3] in 1901 and 1903 mapped a small ice mass on the slopes of Mt. Denson. Grant and Higgins made a geological map of Port Fidalgo in 1908,[4] showing one small non-tidal ice tongue near the head of the fiord. Excepting at Goose Island our own knowledge of Port Fidalgo, as well as the fiords of Port Gravina, and Sheep and Simpson Bays at the south, are based entirely upon observations from the steamers and from launch trips past the mouths of these indentations.

Glaciation Above Sea Level. The fiord walls above sea level suggest the same profound glacial erosion observed in the parts of Prince William Sound already described. There are cirques in the mountains and the fiord walls are rounded and smoothed. Bligh Island (1634 feet high), north of the entrance of Port Fidalgo, has the rounded form that shows it was completely overridden by ice. On the eastern side of Tatitlek Narrows, near Ellamer, the height of glaciation determined by Grant [5] and Capps [6] was 2800 to 3000 feet. Goose Island (320 feet high), south of the entrance of Port Fidalgo, has the rounded form resulting from complete overriding by the glacier. The steepened slope of Knowles Head, between Ports Fidalgo and Gravina, is interpreted as the result of glacial erosion rather than of wave work, which does not at present reach its base, the prominent yellowish landslide upon its face being an indirect recent result of glacial oversteepening.

Submarine Form of Fiord Through Glaciation. Below sea level the soundings also indicate extensive glacial erosion. No soundings have been made in the eastern half of Port Fidalgo beyond Fish Bay; but from this point westward to Prince William Sound the detailed soundings by the Coast Survey [7] show that the outer part of the fiord has a depth of 542 to 714 feet, with a slight westward upgrade in the broad outer part where glacial erosion was less effective because the fiord walls are farther apart. The cross-

[1] Petroff, Ivan, Population, Industries and Resources of Alaska, Tenth Census of the United States, 1880, Vol. VIII, 1884, p. 27.

[2] Schrader, F. C., A Reconnaissance of a Part of Prince William Sound, and the Copper River District, Alaska, in 1898, 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 354 and 380.

[3] U. S. Coast and Geod. Survey, Chart 8519, 1907.

[4] Grant, U. S. and Higgins, D. F., Bull. 443, U. S. Geol. Survey, 1910, Pl. I, facing p. 10, Pl. II, in pocket.

[5] Op. cit., p. 19.

[6] Capps, S. R. and Johnson, B. L., Bull. 542, U. S. Geol. Survey, 1913, p. 89.

[7] U. S. Coast and Geod. Survey, Charts 8519 and 8550.

section of the fiord is broadly U-shaped, the sides descending to a depth of 600 feet in three-eighths of a mile. For a width of $1\frac{1}{4}$ to $1\frac{1}{2}$ miles there is rather flat bottom at a depth of about 600 feet.

Submerged Hanging Valleys. All of the bays tributary to western Port Fidalgo are submerged hanging valleys. Fish Bay (Pl. CXLVI), which is 90 to 96 feet deep at the mouth, hangs at least 500 feet; the bottom of the unnamed cove on the southern side of the fiord opposite Fish Bay is hanging a little less; while Landlocked, Boulder and Bowie Bays, and Snug Corner Cove hang still less above the bottom of the fiord, as would be natural with the greater size of these bays and of the tributary ice tongues which formerly occupied them. Port Fidalgo, 714 feet deep at the mouth, is itself a submerged hanging valley in relation to Prince William Sound, which in a distance of $2\frac{3}{8}$ miles reaches a depth of 1350 feet. This fiord, therefore, hangs over 600 feet above Prince William Sound, where the larger glacier presumably eroded more effectively than the Port Fidalgo Glacier. The channel between Goose Island and Porcupine Point at the southern entrance to Port Fidalgo, which has an average depth of less than 50 feet, and a maximum depth of 150 feet, shows a similar hanging relationship at both the northern and southern ends.

Tatitlek Narrows, with depths of 66 feet or less, hangs 90 feet or more above Boulder Bay, which itself hangs twice that amount above Port Fidalgo. Tatitlek Narrows also hangs above Valdez Arm (Fig. 28) at the northern end. Grant and Higgins state[1] that this strait lies in a belt of weak black slates and has an island of resistant diabase. These narrows constitute a submerged hanging valley of the third order, being discordant in relation to Boulder Bay which hangs above Port Fidalgo which, in turn, hangs above Prince William Sound.

Glacial Deposits. The deposits in the explored portion of this fiord are shown by the Coast Survey chart to be almost uniformly soft or sticky blue mud or fine sand. We have little information about the glacial deposits above sea level along the sides of Port Fidalgo and the tributary valleys. There are small terraces of outwash gravel near the entrance of the fiord. Bligh Island has "gravel deposits containing exotic materials," as Schrader and Spencer noted in 1900,[2] and there is also till with foreign materials on Goose Island, as we observed in 1910.

Port Gravina. Port Gravina, southeast of Port Fidalgo, is of smaller size, having a length varying from 13 to 16 miles, as shown on different maps. It flares open rather widely at the mouth, where it is half as wide again as Port Fidalgo. This fiord extends back into the Chugach Mountains, with peaks on either side rising to about the same heights as those around Port Fidalgo and with steep walls, those on the southern side rising to 2162 feet within half a mile of the shore.

Most if not all of the glaciers about the head of Port Gravina are small and only a few of them have even been indicated upon maps, two small ones, sketched from a distance, being shown on the Grant and Higgins map. None of them has been studied by any persons exploring Prince William Sound, and we observed them only from a distance near the mouth of the fiord. Glacial erosion and glacial deposits have not been investigated as yet, and there are no soundings at Port Gravina, so that we lack all specific

[1] Op. cit., p. 50.

[2] Schrader, F. C. and Spencer, A. C., Geology and Mineral Resources of a Portion of the Copper River District, Alaska, House Doc. 546, 56th Congress, 2nd Session, Washington, 1901, p. 81.

information about conditions here except in connection with certain gravel deposits near the entrance.

Orca Bay and Orca Inlet—Topography. These two inlets border Hawkins Island on the northern and southern sides, Orca Bay, which has also been called Cordova Bay, being tributary to Prince William Sound just south of Port Gravina, while Orca Inlet opens into the Pacific Ocean south of Hawkins and east of Hinchinbrook Island. The two inlets unite east of Salmo Point on the eastern end of Hawkins Island, forming a fiord 6½ miles in length with a width of a little less than 1½ miles, at the upper end.

Orca Bay extends westward from Salmo Point for at least 18 miles, increasing greatly in width, so that between Gravina Point and Hawkins Island Cutoff it has a breadth of 7½ miles. Simpson and Sheep Bays are its tributaries on the northern side, and on the southern side are Windy Bay, Cedar Bay, and numerous other small indentations, such as Hawkins Island Cutoff and Canoe Passage, which connect this part of Prince William Sound with Orca Inlet and the Pacific Ocean.

Orca Inlet extends southwestward for about 16 miles, not including the portion east of Salmo Point nor that leading to Hawkins Island Cutoff, increasing in width to 3 miles. Odiak Slough enters Orca Inlet at the town of Cordova. A great contrast between Orca Inlet and Orca Bay is that the southwestern two-thirds of the former is so shallow that it is navigable only by small boats, while the latter is a deep fiord, traversed by ships on their way to Cordova.

The shores of both Orca Inlet and Orca Bay are high and steep, as in the case of the other fiords of western Prince William Sound, though in each case the outer, western portions are much lower than the inner portions. The eastern half of each of these inlets has walls rising to elevations of 2000 to 3000 feet or more.

Size of Glaciers. The largest glaciers at present known to be tributary to any of the fiords of eastern Prince William Sound, excepting Port Valdez, descend from the snow-capped mountains east of the end of Orca Inlet. The greater size of these ice tongues is perhaps due to the fact that this part of the Chugach Mountains is higher than that near the heads of Ports Fidalgo and Gravina, or perhaps to the heavier precipitation upon this portion of the mountains nearest the ocean. The largest of these glaciers, none of which reach tidewater, is Shephard Glacier.

Shephard and Adjacent Glaciers. Shephard Glacier which has not been visited or studied by any geologist, though roughly mapped from a distance by Grant and Higgins,[1] (Pl. XCIII) comes from snowfields a considerable distance back in the mountains. It flows southwestward from the portion of the Chugach Mountains east of the head of Orca Inlet, and has an unusually irregular shape. The upper part has a width of about 1½ miles, and sends two distributary ice tongues northward about a mile into the valley which continues northeastward from the head of Orca Inlet. Below these distributaries the main glacier continues southwestward, for about 3½ miles, with an average width of ¾ of a mile, the terminus being 2½ miles east of the head of Orca Inlet. From this terminus of Shephard Glacier a stream flows to Eyak Lake, through a valley which parallels Orca Inlet. Another glacial distributary east of the main ice tongue flows southward for about 1½ miles and gives rise to Ibeck Creek which joins the stream of Scott Glacier which flows into the Pacific Ocean east of the valley of Eyak Lake. Shephard Glacier, therefore, has (a) one small distributary whose glacial stream flows directly into the Pacific Ocean;

[1] Bull. 443, U. S. Geol. Survey, 1910, Pls. I and II; *Ibid.*, Bull. 379, 1909, Pl. IV, facing p. 88.

(b) a main ice tongue whose stream flows into Eyak Lake, which drains by Mountain Slough over the Copper River delta to the Pacific Ocean; and (c) two smaller distributaries, which send streams directly to the head of Orca Inlet.

Since we have not visited Shephard Glacier we lack further information about it. There are two unnamed ice tongues to the north, each of which is over half a mile in width. Their streams join the one from Shephard Glacier which flows to the head of Orca Inlet.

Glaciers of Sheep and Simpson Bays. The mountains at the head of Sheep and Simpson Bays rise precipitously to a considerable height and have small glaciers, none of which have been mapped or observed from nearer than the entrance to these bays. The fiord walls are very steep, the northwestern side of Sheep Bay rising to an elevation of 2520 feet within a mile of sea level. There are many large cirques in these mountains, due to extensive former local glaciation. Within Sheep Bay the depth of water varies from 144 to 300 feet with a shallower inner portion studded with islands, suggesting that there has never been a very great glacier extending from the head of this inlet. The upper part of Simpson Bay is also irregular, with large peninsulas.

Glacial Erosion in Orca Bay and Orca Inlet. These arms of the sea are typical fiords and have the usual characteristics of marked modification by glacial erosion. The higher mountains at the head of Orca Inlet and Simpson and Sheep Bays, have large areas of rounded, bare rock and many cirques and hanging valleys. Some of the lower slopes of the fiord walls are much oversteepened by glacial erosion, the southeastern side of Orca Inlet, between Orca cannery and Cordova, for instance, rising to 1600 feet within a half mile of sea level, but only 898 feet in the next half mile. All of the fiord walls do not rise so steeply, however, and parts of outer Orca Bay have rather low, sloping shores. In Orca Inlet the smoothing by the former ice tongue extends up to 2300 feet, and north of Orca Bay to 2500 feet, as estimated by Grant and Higgins.[1] In the town of Cordova observations of the direction of striae, by the junior author in 1910, make it clear that the ice of one of the expanded lobes of Shephard Glacier advanced down the valley of Eyak Lake, and spilled over into Orca Inlet. This resulted in notable shortening of the spur upon which Cordova is built.

Glacial erosion below sea level is shown very well by the detailed soundings on the Coast Survey charts.[2] Orca Bay has a depth of 198 feet at Salmo Point, and deepens rapidly to 744 feet opposite Simpson Bay, 7½ miles from Salmo Point; but this is an exceptional depth, for in the next 11 miles, westward to the mouth, the depth is only 400 to 500 feet. The portion opposite the mouth of Simpson Bay may be a basin due to glacial scooping. Near its outer edge there are rock reefs including Hanks Island, Gatherer Rock, and a series of shoals which extend nearly halfway across Orca Bay. Sheep Bay, which is as wide at the mouth as Orca Bay at this point, does not appear to have a hanging relationship to Orca Bay. Simpson Bay occupies a submerged hanging valley, being 120 to 336 feet deep at the entrance, while just outside the lip of the hanging valley, the depth is 516 feet, a discordance of at least 180 feet. Windy Bay, on the southern side of Orca Bay, is a submerged hanging valley, having a depth of 114 to 174 feet at the entrance and hanging 546 feet above the bed of Orca Bay, which is here 720 feet deep. Both Canoe Passage and Hawkins Island Cutoff are apparently hanging valleys, but there is an unknown thickness of deposit in each.

[1] Grant, U. S. and Higgins, D. F., Bull. 443, U. S. Geol. Survey 1910, p. 19.

[2] U. S. Coast and Geod. Survey, Charts 8520 and 8550.

The amount of glacial erosion below sea level in most of Orca Inlet is entirely concealed by deposits. The part of Orca Inlet from Cordova to the head has depths of from 90 to 228 feet east of Salmo Point, but is much shallower from that point to Cordova, with depths of from 6 to 30 or 40 feet and with many shoals and a few navigable channels in which the maximum depth of water is 90 feet. Southwestward from Cordova the depth is generally 12 feet or less, excepting for one or two small discontinuous channels. Thus it is evident that Orca Inlet differs from all the other fiords in Prince William Sound, and indeed from most Alaskan fiords, in being shallow. It is a fiord nevertheless but one very much modified by deposition.

Glacial Deposits on the Land. Glacial deposits around Orca Bay and Orca Inlet are inconspicuous, being thin or absent because of the steepness of the fiord walls. Unconsolidated till with striated, angular bowlders may be seen in a number of cliffs along the coast, near Cordova, and in the town. There are also small patches of outwash gravel, including several near the town of Cordova. The largest area where these are developed, so far as we know, is a large valley train, described to us by prospectors as occupying the valley which leads northeastward, from the head of Orca Inlet to the northern margin of Shephard Glacier and the adjacent ice tongues. There are small terraces of outwash gravel on the mainland north of Orca Bay, and on the northern side of Hawkins Island at the narrowest portion of the Salmo Point peninsula.

Submerged Glacial Deposits. In these fiords the glacial deposits below sea level are the most extensive and we shall describe in order (1) the bottom deposits of Orca Bay, which do not seem to form submerged moraines but mantle the bottom of the fiord, (2) the deposits which have filled up the larger part of Orca Inlet, and (3) the Middle Ground Shoal of Orca Bay, which is a tidal delta built by glacial marine deposits from Orca Inlet.

In the fiords of the northern and western sides of Prince William Sound, the deposits below sea level are imperfectly described, because, in connection with the soundings which we made, no samples of bottom material were collected. Accordingly, we have thus far discussed only glacial deposits on the sea bottom where they stood up as moraines. In Orca Bay and Orca Inlet the detailed Coast Survey charts furnish some indication of bottom material as well as depths. We can, therefore, describe, in a general way, the bottom deposits in Orca Bay, which seem to have been made by a greater glacier filling the whole fiord and which probably retreated eastward into the mountains northeast of Orca Inlet and the mountains northeast of the heads of Sheep and Simpson Bays. These deposits are all to be classed as glacial, for no sediment-laden streams of any size now enter Orca Bay, and the former rivers, whose deposits partly mantle the material dropped by the retreating glaciers and by floating icebergs, were all glacial streams.

In the eastern part of Orca Bay, from Bomb Point, to the Salmo Point Narrows the chart indicates rocky bottom and sand, a condition not found elsewhere in the fiord excepting close to the shore. This suggests a bottom veneer of morainic deposits. There are small patches of mud in mid-fiord.

From this point westward the fiord has a bottom of sticky or soft mud, evidently brought in by tidal currents, with hard material only rarely, and with rocky bottom only near shore.

The shallow western two-thirds of Orca Inlet has been so modified by filling that most of it is unnavigable, the filling having been largely done by tidal currents, which are

still engaged in the same work. These deposits may be spoken of as glacial deposits because they were formerly and are still being supplied almost entirely by streams from melting glaciers farther east, notably the Scott, Sheridan, and Martin River Glaciers, and smaller ice tongues along the mountain front, and the Miles, Childs, Goodwin, Allen, Heney, and other glaciers of the Copper River valley. The Copper River, a great overburdened glacial stream, has built an enormous delta, coalescing with the deltas of Eyak and Martin Rivers. Much of the finer material is carried out beyond this delta and is drifted westward along shore and, since Orca Inlet is the first great opening in the coast, a large amount of the sediment has been driven into it, so nearly filling it that only narrow, shallow channels are occupied by water at low tide.

The third deposit of glacial origin below sea level, is Middle Ground Shoal, a triangular area projecting into Orca Bay from Hawkins Island Cutoff, and interpreted as a tidal delta. It is composed of fine sand and clay similar to that filling Orca Inlet and derived from the same source. This sediment is part of the material that was drifted into Orca Inlet in suspension, and swept through Hawkins Island Cutoff, coming to rest where the tidal currents are checked in the broad outer portion of Orca Bay.

This deltoid shoal, 5 miles wide near the land, extends northward 3½ miles into Orca Bay. It is covered at mean low tide by water only 1½ to 15 feet deep, and has narrow distributary channels 21 to 30 feet in depth. At the front and edges of this tidal delta the depth increases rapidly, reaching 156 feet in a little over a quarter of a mile and 300 or 400 feet a little further out. The triangular deposit has, therefore, been built in deep water and is still being extended. Its margins slope 7½° to 10°. The channel in Hawkins Island Cutoff and the distributary channels on the tidal delta are kept open by the tide, which moves both ways through the strait.

CHAPTER XIX

GLACIERS NEAR THE COPPER RIVER DELTA

General Relationships

The glaciers described in this chapter include:—(1) the Scott, Sheridan, and Sherman Glaciers, between the town of Cordova and the Copper River; (2) the Goodwin, Fickett, Saddlebag, and McPherson Glaciers, a group of ice tongues on each side of Copper River between its delta and the Childs and Miles Glaciers; and (3) Johnson and Martin River Glaciers, east of the Copper River delta at the base of the Chugach Mountains. Some of these glaciers drain into the Copper River but the streams from three of the largest ice tongues pursue independent courses to the sea, forming outwash gravel plains and deltas, which coalesce with the great delta of the Copper River.

Glaciers West of Copper River

Scott Glacier—General Description. There are several small glaciers in cirques along the mountain front, but Scott Glacier is the first large ice tongue seen from the Copper River and Northwestern Railway between Cordova and the Copper River. It occupies a valley immediately east of the mountain ridge in which Eyak Lake lies and has a known length of nearly six miles and a width of about a mile near the terminus where there is a lake half a mile long.

The surface of Scott Glacier does not seem from a distance to be much crevassed and there are several medial moraines. We lack further information about it, however, not having seen it from nearer than the railway, a distance of 7 miles. No one interested in glaciers seems to have visited or described it and the only general representations of it upon maps were made by Allen [1] in 1885, Abercrombie [2] in 1898, and Grant and Higgins [3] in 1908.

Outwash Plain. Between the glacier and the railway is an extensive outwash gravel plain, parts of which are still being built up by Eyak River and by the heavily-laden stream from Scott Glacier, which is reinforced by a stream from one of the eastern lobes of Shephard Glacier. Where these streams are actively building, the plain is barren, but the portions of it over which streams have not recently worked are heavily forested, and some parts between the railway and the ocean are extensive grassy marshes.

Sheridan Glacier—Previous Descriptions. Sheridan Glacier is a conspicuous ice tongue,

[1] Allen, Henry T., Report of a Military Reconnoissance in Alaska, Narratives of Explorations in Alaska, Senate Rept. 1023, 56th Congress, 1st Session, Washington, 1900, map facing p. 434.

[2] Abercrombie, W. R., War Dept., Adj.-Gen. Office, No. XXV, 1899, map in pocket. Abercrombie doubtless made maps in 1884 of this and some of the other glaciers described in this chapter, but they were not published in his report and are not now on file at the War Department.

[3] Grant, U. S. and Higgins, D. F., Bull. 379, U. S. Geol. Survey, 1909, Pl. IV, facing p. 88; *Ibid.*, Bull. 443, 1910, Pls. I and II.

about 15 miles east of Cordova (Pl. XCIII), terminating a little over 2 miles north of the railway. It has been briefly mentioned and sketched upon maps by Abercrombie [1] in 1884 and 1898, Seton Karr[2] in 1886, and Witherspoon[3] in 1900 and 1911. Our own observations are based upon two days work in 1910 by the junior author and general views from the railway in 1909 and 1911.

Valley Portion. The glacier has a well-defined valley portion, rising in snowfields 5000 or 6000 feet above sea level and supplied by at least two main tributaries. Four or five cascading ice tongues reach the glacier, and there are two or three hanging glaciers which no longer supply ice to it. The valley tongue of Sheridan Glacier is about a mile wide and at least 5½ miles long. It is clean and little crevassed and has one weak medial moraine near the eastern margin.

Bulb-Shaped Terminus. Beyond the mountain face the glacier expands into a bulb-shaped terminus, sloping in all directions from the mouth of the valley. This bulb, is over 4 miles long and 3 miles wide, terminating about 150 feet above sea level. Most of its surface is without moraine and it was little crevassed in 1910. The terminus, however, bears a narrow band of ablation moraine, broadest near the western edge. Close to the front of Sheridan Glacier bulb is a low, narrow terminal moraine, covered with a thick growth of alder, cottonwood and some conifers.

Seton Karr's drawing and description of the glacier in 1886 [4] makes it clear that the expansion of the bulb took place more than 25 years ago and that there has been little recession since that time. The presence of thick mature forest up to the very edge of the bulb indicates that the glacier has not been more extensive for a score of years, perhaps for a century. A comparison of photographs shows that there was no significant change between 1908 and 1910, with the exception of a slight thinning by ablation; and there was no visible change from 1910 to 1911. At the very front of the glacier, and east of the middle, is a great rock knob to the summit of which the ice rises, but without passing over it or covering the southern side. This knob gives a convenient basis for determining future oscillations of the front of Sheridan Glacier in the way of moderate advance or of a continuation of the very slight recession which has characterized the recent history of the glacier.

Outwash Plain. Between the glacier and the sea lies an extensive outwash plain. The stream from the eastern edge of Sheridan Glacier, probably also carrying the drainage from Sherman Glacier, is building an extensive alluvial fan which is separated from the main outwash gravel plain by a line of forest-covered hills near Mile 17, several hundred feet high. A number of other knobs and spurs rise like islands through the outwash gravels. East of this alluvial fan, the railway crosses several low ridges, some of till, some of rounded gravels, and a few of rock, and all densely forested. It is clear that these unconsolidated ridges represent remnants of a terminal moraine of an eastward

[1] Abercrombie, W. R., Report of a Supplementary Expedition into the Copper River Valley, Alaska. Narratives of Explorations in Alaska, Senate Rept. 1023, 56th Congress, 1st Session, Washington, 1900, p. 384; *Ibid.*, War Dept., Adj.-Gen. Office, No. XXV, 1899, map in pocket.

[2] Seton Karr, H. W., Shores and Alps of Alaska, London, 1887, pp. 169–171.

[3] Witherspoon, D. C., Pl. II in House Doc. 546, 56th Congress, 2nd Session, 1901; also in Bull. 374, U. S. Geol. Survey, 1909, Pl. I; for detailed map made in 1911, see Brooks, A. H. and others, Railway Routes in Alaska, House Doc. 1346, Part 2, 62nd Congress, 3rd Session, Washington, 1913, Plate 5.

[4] Op. cit., p. 171.

expansion of Sheridan Glacier, rather than a westward expansion of the former Copper River Glacier.

A considerable portion of the main outwash gravel plain, north of the railway, and of the alluvial fan to the east of it, is forested; but where the present streams flow there are barren areas and extensive groves of dead trees, the lower portions of their trunks being deeply buried in gravel. The killing of mature spruces by alluviation was still in progress in 1910. South of the railway much of the outwash gravel plain is covered with swamp grass because it is too wet for tree growth, while still nearer the Pacific there are broad areas of salt marsh and mud flat.

Sherman Glacier—General Description. Sherman Glacier,[1] occupying the valley east of Sheridan Glacier (Pl. XCIII), has its main source 3200 feet above sea level in the mountains immediately west of the Copper River. It flows southwestward with a length of over 7½ miles, and a width of 1½ miles, terminating entirely within its mountain valley.

Relation to Sheridan Glacier. Sheridan Glacier bulb completely overlaps the mouth of this valley, but Sherman Glacier supplies none of the ice of the present bulb. The early maps all show Sheridan and Sherman Glaciers uniting and forming a piedmont ice mass; but this was discovered to be an error by the junior author in 1910 and is correctly represented on Witherspoon's revised map of 1911. A view from the fan of outwash gravels makes it clear that Sherman Glacier is not a tributary, but an independent glacier, though the two may be in actual contact in a short portion of the area not visible from our point of observation. A rock hill rises through the Sheridan ice in one portion of the margin. The two glaciers may have fed one piedmont area at a more expanded stage of glaciation, but it is thought possible that both ice tongues had receded into their mountain valleys before the Sheridan readvanced and formed the present ice bulb.

Tributaries. In addition to the ice supply from the 3200 foot divide, Sherman Glacier is fed by three or more steeply cascading tributaries, descending the slopes of Mt. Murchison [2] from the north. Near the terminus there are two prominent hanging valleys out of the mouths of which cascading glaciers extend, though they do not now reach down to the surface of the main glacier. There are several weak medial moraines, but the surface of the glacier is very clean and little crevassed. Sherman Glacier though wider than Sheridan Glacier in its mountain valley, and with more medial moraines, and presumably more tributaries, is apparently at present the weaker of the two; for, as indicated above, it seems not to have participated in the advance which carried the bulb of the Sheridan out past the Sherman terminus.

Through Glacier. Rising on a broad snow divide, from which a smaller ice tongue extends eastward to the Copper River valley, Sherman Glacier is an excellent illustration of the type of through glacier, for which Hobbs [3] has proposed another name, which to us seems much less suggestive, figuring this glacier as an example. The map upon which he based his figure [4] is an older one upon which the topographer transferred the name of Sheridan Glacier to Sherman Glacier, though Abercrombie, who named the former, applied it to the westernmost of these two ice tongues.

[1] Named in 1910 for General W. T. Sherman.

[2] Named in 1910 for S. Murchison, one of the engineers who was prominent in the building of the Copper River and Northwestern Railway.

[3] Hobbs, W. H., Characteristics of Existing Glaciers, New York, 1911, pp. 44–5.

[4] Op. cit., Fig. 16, p. 45.

Glaciers in the Lower Copper River Valley

Fickett Glacier.[1] From the 3200 foot divide of the through glacier of which Sherman Glacier is the western end, an ice tongue descends eastward toward the Copper River valley. This is Fickett Glacier,[2] an ice tongue nearly 4 miles long and from half a mile to a mile wide. It descends over 2500 feet from snowfield to terminus and has one steeply-cascading tributary from a cirque on the southern side. It is just barely visible from the Copper River railway and we have not approached it closely.

Saddlebag Glacier. Saddlebag Glacier is a smaller, double-ended through glacier, south of Fickett Glacier. Its two members rise on a 3000 foot divide, the shorter, eastern portion having a length of 1½ miles and terminating in the Copper River valley about 900 feet above sea level, while the larger, southern portion is 3½ miles long and descends to within 200 feet of the level of the sea. This southern portion of the glacier, which occupies the deep valley on the seaward face of the Chugach Mountains, between McKinley Lake and Copper River, is a third of a mile wide and is fed not only by the ice tongue from the through divide, but by a tributary from a cirque on the west. A photograph by H. P. Ritter of the U. S. Coast and Geodetic Survey shows that in 1899 it was clean and little crevassed, and had a narrow barren zone in front, and a valley train of outwash gravels, extending southward to Copper River. Its valley walls, seen just before crossing the Copper River on the first steel bridge, have been rendered very precipitous by glacial erosion. There are also small glaciers north of McKinley Lake.[3]

Goodwin Glacier. Goodwin Glacier, which is immediately south of Childs Glacier, has a known length of nearly 7 miles and a width in the mountain valley of three-fourths of a mile. On emerging into Copper River valley it expands into a piedmont bulb between two and three miles in width. It rises in cirques and snowfields 5000 to 6000 feet above sea level on the slopes of Mt. Murchison, and slopes gradually, at an angle of 7° in the lower portion, but with no visible hanging valley relationships. The number of tributaries is not known, but there are two conspicuous ones from the south, which, however, give rise to no medial moraines. The terminus, rising about 400 feet above the river, does not form an ice cliff in Copper River as the Childs Glacier does, but the ice extends almost to the water's edge being separated only by a crescentic area of terminal moraine and outwash, most of which is thickly covered with alders and with scattered cottonwoods. Inside this forest belt is a barren zone of moraine-covered ice, also crescentic in form, and extending up to the clear ice of the valley glacier, where in 1910 there were a few reddish, moraine-covered ice cones and some grayish ablation moraine.

Goodwin Glacier is a conspicuous object from portions of the Copper River and Northwestern Railway. It was named by Abercrombie,[4] who in 1884 camped upon a barren part of the outwash deposit, and is shown, in a general way, on his 1898 map. This map and the allusion by Seton Karr[5] to Goodwin and Childs Glaciers as ice tongues which spread out in 1886 "with beautiful fan-like shape to the river level," make it clear that the advance and expansion of the Goodwin bulb took place over 25 years ago. The

[1] Named in 1910 for Fred W. Fickett, the soldier who accompanied Lieutenant Allen up the Copper River in 1885.

[2] Not Goodwin Glacier as an early map suggested.

[3] See map by D. C. Witherspoon, Bull. 542, U. S. Geol. Survey, 1913, Pl. III, facing p. 78.

[4] Abercrombie, W. R., Supplementary Expedition into the Copper River Valley, Alaska, Narratives of Explorations in Alaska, Senate Rept. 1023, 56th Congress, 1st Session, Washington, 1900, p. 387.

[5] Seton Karr, H. W., Shores and Alps of Alaska, London, 1887, p. 170.

forest at the very edge of the glacier proves that the glacier has not been notably larger for a much greater period.

McPherson Glacier. McPherson Glacier[1] is a moderate-sized ice tongue on the eastern side of Copper River south of Miles Glacier. The main ice tongue has a length of over 4 miles, and a width at the terminus of about a quarter of a mile. It is fed by two tributaries from the southeast, each about 3 miles in length, the northernmost heading on a 6200 foot divide from which a tributary also flows northward to the Miles Glacier. This ice tongue is not known to have oscillated significantly in recent years. It was retreating when visited by the junior author in 1910.

In its lower portion McPherson Glacier emerges from a hanging valley, and the terminus of the glacier cascades 800 to 1000 feet over the lip of this hanging valley, ending at the base of a steep rock slope at an elevation of 310 feet above sea level. It is a conspicuous object from the railway, because of the clean, crevassed condition of this terminal cascade. That the cascading portion of the glacier is thin, is proved by the emergence of a rock ledge about half way up the slope; and it is as little crevassed as a glacier could be on so steep a slope. This terminal ice cascade occupies only the southern half of the lip of the hanging valley, and that it has recently diminished several hundred feet is proved by a barren zone on the northern side. For many years the ice has not covered the remainder of the northern half of the valley lip, for there are dense thickets of willow and alder and scattered spruce trees upon it. There is no further evidence as to recent changes in this glacier.

At a distance of about 100 yards in front of McPherson Glacier, striated, rounded rock ledges constrict the valley, and a short distance farther west another rock ledge from the north projects part way across the valley. On approaching McPherson Glacier from the railway, these ledges give one the impression that they are large terminal moraines, but although morainic material has accumulated on them, there are no independent moraines here, though there is a low terminal moraine nearer the glacier.

Sheep Creek, the stream from McPherson Glacier, has built an extensive outwash gravel plain, above the level of which are forested terraces of older outwash. The plain broadens westward and coalesces with the great outwash gravel plain of the Copper River.

Glaciers East of Copper River Delta

Johnson Glacier. Johnson Glacier[2] (Pl. XCIII), south of McPherson Glacier, is fed by four long narrow tributaries, the main ice stream having a length of 5½ miles, and a width of about half a mile near the terminus which is less than 200 feet above sea level. It is over 8 miles from the end of the main glacier to the head of the longest of the tributaries. All these tributaries rise in snowfields 4000 to 5000 feet above sea level, the longest one on the north heading on a divide which also sends a tributary northward to Miles Glacier. Each of the tributaries cascades steeply down the mountain slopes.

[1] Named in 1910 for J. L. McPherson of Seattle who made a railway survey up the Copper River valley during the winter of 1906.

[2] Named in 1910 for Albert O. Johnson, one of the bridge engineers of the Copper River and Northwestern Railway, who made many instrumental observations upon the glaciers and glacial streams between 1908 and 1910.

This glacier was mapped by Witherspoon in 1900,[1] and is shown with different form in the revised map published in 1909.[2] It has been seen by the National Geographic Society's expedition only from a distance, and its chief importance in our studies comes in connection with its contribution of outwash to the large valley trains of the Copper and Martin Rivers.

Martin River Glacier. Martin River Glacier terminates about 20 miles east of Copper River, and is separated from the sea by the foothill region of the Controller Bay coal field. It is 3 to 5 miles wide, over 25 miles long, and terminates less than 350 feet above sea level. It is larger than Miles Glacier, much larger than Sheridan Glacier, and is the most extensive valley glacier in the Copper River portion of the Chugach Mountains.[3] The covering of ablation moraine on the terminus renders the glacier relatively inconspicuous from the Copper River railway, and it would not be noted here were it not for the extensive outwash gravel plain built up by its glacial streams. These coalesce with the Copper River delta on the west, so that Copper River has been pushed to the extreme western edge of its valley.

Glaciation Near Copper River Delta

The varied and interesting features produced by glacial erosion and glacial deposition near the ice tongues described in this chapter are so involved with the other glaciers to be described and with general conditions of glaciation of the Copper River valley, that their treatment is postponed to Chapter XXIII.

[1] Plate II, House Document 546, 56th Congress, 2nd Session, The Geology and Mineral Resources of a Portion of the Copper River District, Alaska, by F. C. Schrader and A. C. Spencer, Washington, 1901. See another arrangement of glaciers here, in the map published as Pl. XII, Bull. 284, U. S. Geol. Survey, 1906.

[2] Pl. I, Bull. 374, U. S. Geol. Survey, 1909; also Chitina Quadrangle.

[3] This glacier has been described by G. C. Martin. See Bull. 335, U. S. Geol. Survey, 1908, pp. 16, 48–52, 56, 64–65 and Plates I, II and III. The junior author of this book visited portions of the glacier in 1904.

CHAPTER XX

CHILDS GLACIER

General Description. Childs Glacier (Map 9), entering Copper River valley from the west, just north of Goodwin Glacier, at the head of the Copper River delta, ends in the river in a vertical ice cliff from 200 to 300 feet high. This ice cliff is all that one sees of the glacier from the railway bridge or from the river bank near Miles Glacier railway station. The ice cliff is so imposing that it has led some visitors to believe that Childs is larger than Miles Glacier, which is farther away, when as a matter of fact Childs Glacier is less than half as large.

The ice tongue is about ten or twelve miles long, rising in cirques and on snowy divides six or eight thousand feet above sea level (Pl. CXLVII), and is fed by at least five tributaries. The width of the valley glacier (Fig. 58) varies from a little over one to a little less than two miles, the narrowest part being near the mouth of the mountain valley. East of this constriction (Fig. 57) the glacier expands into a small imperfect bulb in the Copper River valley, increasing in width from a mile and an eighth to over three miles in a distance of less than two miles. The low, sloping northern and southern margins of the glacier end on the land, being faced on each side by an alluvial fan; but in the middle the glacier terminates in Copper River, whose depth in front of the ice cliff is from 10 to 20 feet, and whose undercutting action has produced precipitous ice cliffs for a distance of over two and a half miles.

The glacier surface is, in the main, clean and white, excepting along the terminal margins, and along the narrow lateral moraines extending up the glacier from these. There are four or five weak medial moraines, coming from the junction of tributaries in the mountains and curving irregularly with the flow of the ice in the expanded bulb. These medial moraines are not conspicuous features on the terminal cliff.

From its terminus to the snowfields, the glacier is severely crevassed, excepting in the northern and southern margins of the bulb where, up to 1909, the crevassing was so slight that the ablation moraine completely mantled the ice and even supported a few shrubs and annual plants. From the mouth of the mountain valley to the margin of the bulb, the slope of the lower portion of the glacier is 382 feet to the mile, an elevation of 1500 feet above sea level being reached less than three miles from the river. Above this the slope may be steeper, though the glacier maintains a moderately low grade as far as seen. There are, however, a number of cascading tributaries.

If Childs Glacier valley hangs in relation to the main Copper River valley it is completely masked by the glacial deposits. The walls of the glacier valley (Fig. 57) rise steeply to elevations of from three to six thousand feet in a short distance, as is well shown north of the terminus, where a spur on the slopes of Mt. O'Neel[1] (Pl. CLIII)

[1] A 6400 foot peak named in 1910 for Mr. A. C. O'Neel, chief bridge engineer of the Copper River and Northwestern Railway, who built the steel bridge across the Copper River near the base of this mountain.

reaches an elevation of 3857 feet, with a slope which rises at the rate of over 4000 feet to the mile. Upon this steepened slope there are several small, steeply-cascading glaciers which are no longer connected with the main ice stream.

Childs Glacier terminates close to the railway, and is seen annually by many travelers, who view it from the train at Miles Glacier station or walk down the path along the river bank west of the railway bridge. Its magnitude may be better appreciated by seeing how the capitol in Washington (Pl. CXLVIII, B) would appear in comparison with the precipitous cliff that rises above the river, and how its ice stream (Fig. 60) compares in width with the Nisqually Glacier on Mt. Rainier, one of the largest ice

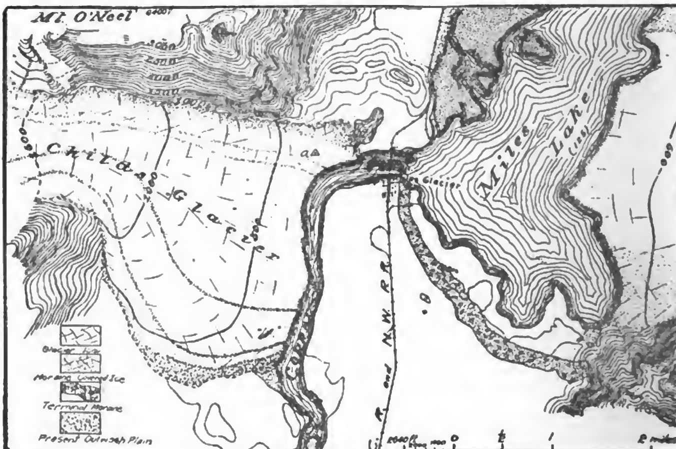


Fig. 57. Childs Glacier in 1910.

tongues in the United States, itself far larger than the greatest of the ice masses seen along the Canadian Pacific Railway in the Selkirks and Canadian Rockies.

Observations before the Building of the Railway. It is not known when or by whom the earliest observation of Childs Glacier was made, but as early as 1850 the Russian geologist Grewingk recorded [1] that the Copper River crosses mountains, "the ravines of which are filled with perpetual ice; and, where the river breaks through, the ice masses are undercut, torn away, and precipitated into the river with terrific crashing." This probably applies to Childs or Miles Glacier, or both.

C. G. Holt ascended Copper River in 1882 but we have no information regarding what he saw of the glaciers.

[1] Grewingk, C., Beitrag zur Kenntniss der Orographischen und Geognostischen Beschaffenheit der Nord-west-Küste Amerikas mit den Anliegenden Inseln, Verhandl. Russ. Kaiserl. Mineral. Gesell. zu St. Petersburg, 1848 und 1849, pp. 34–5.

In July, 1884, Abercrombie ascended the Copper River, taking several photographs of Childs Glacier, experiencing much difficulty because of the icebergs in the river, and noting the speed of the current, the large bowlders carried, and the loud reports from the discharge of icebergs. He estimated that the glacier delivered 8,160,768,000 pounds of ice yearly to the river.[1] On his way down the river, in September, he ran the rapids in front of Childs Glacier.

On March 31st and April 1st, 1885, Allen was passing Childs Glacier, where he noted that the river was confined to one channel about 125 yards wide, east of which was a deposit "considerably elevated above the river bed, and overgrown with small timber, which was so thick as to be a great impediment to the movement of our sleds." He also mentions "the condition of the ice in front of this (deposit) forbidding an attempt (to travel) along the river." Allen's report is illustrated by several woodcuts of Childs Glacier from photographs taken by Abercrombie the year before,[2] which show that the glacier expanded in a clean bulb of white ice in 1884, much as in 1909, having a terminal slope of about the same steepness.

In August, 1891, Hayes and Schwatka descended Copper River, and Hayes made a sketch map showing the Childs and other glaciers and "passed the Childs Glacier, running within a stone's throw of the lofty wall of ice,"[3] which suggests lack of great activity of the ice front at that time.

In October, 1898, Abercrombie descended Copper River, observing great changes in the faces of both Childs and Miles Glaciers since his visit in 1884. Childs Glacier then no longer ended in the river, having "a beach some 500 or 600 yards in breadth between it and the water, but adding a succession of very boisterous rapids,"[4] including "200 or 300 yards of very nasty, rapid water," below which "the river is deflected to the left (west) by a mass of terminal moraine." Corporal Koehler, who accompanied Abercrombie on this journey past Childs Glacier, stated that "this glacier is dry, because it does not throw any ice. There are some pretty high swells along this glacier, and as the current ran about 8 miles an hour, the boat was sucked with the speed of an arrow."

Guide Rafferty furnishes evidence either that Childs Glacier had retreated westward during the summer of 1898, or, what is more likely, that Abercrombie went down the river at a stage of very low water, for he stated that in early July of the same year "the next danger was from the swell caused by the ice breaking off from the Childs Glacier, which sometimes created such waves as to land a loaded boat 150 feet high and dry on the shore. The current swings directly toward the glacier on making the turn, and it required all the strength at hand to keep from being carried with it. The river was filled with floating ice, some pieces being almost the size of a freight car."

[1] Abercrombie, W. R., Supplementary Expedition into the Copper River Valley, Alaska, 1884, Narratives of Explorations in Alaska, Senate Rept. 1023, 56th Congress, 1st Session, Washington, 1900, pp. 386–390.

[2] Allen, H. T., Expedition to the Copper, Tanana and Koyukuk Rivers, Washington, 1887, p. 42, Figs. 3 and 4 and Map 2; also in Narratives of Explorations in Alaska, Washington, 1900, pp. 424–5.

[3] Hayes, C. W., Expedition Through the Yukon District, Nat. Geog. Mag., Vol. IV, 1892, p. 126.

[4] Abercrombie, W. R., Reports of Explorations in the Territory of Alaska, 1898, War Dept., Adj.-Gen. Office, No. XXV, Washington, 1899, pp. 315, 322.

Koehler, R. A., Same, p. 431.

Rafferty, J. J., Same, p. 450.

These accounts by Messrs. Abercrombie, Koehler, and Rafferty are also in the Narratives of Explorations in Alaska, Washington, 1900.

H. L. Wilson, Jr., who was passing Childs Glacier in the spring of 1898, corroborates the testimony of the guide Rafferty that the glacier then reached the river. His description is as follows:

"On June 12 we were just below the Childs Glacier, on the right bank of the river. At 4:30 a. m. we commenced to line our boats past the glacier. Here the river is composed of one stream and, in places, runs very swift. It is about one-fourth of a mile wide. The glacier at this time of the year is along the water's edge. It is about a mile and a half long and 150 feet high. It is supposed to be moving forward. The opposite side or bank is a long rocky beach. When the glacier "dumps" it throws a swell varying in size to the amount of ice that falls. We sent our first two boats up on the rocky beach, and another was swamped at the water's edge. . . . It took a swell about three minutes to reach us after the ice had fallen, the second or third swell being the larger and stronger.

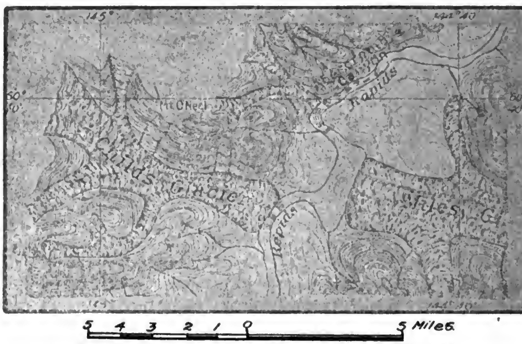


FIG. 58. MILES, CHILDS, AND GRINNELL GLACIERS IN 1900.
Contour interval 200 feet (After D. C. Witherspoon, U. S. Geol. Survey).

On the third trip one of our boats was thrown so high and dry that the men on the lines, nine in number, were in water over their heads and were obliged to cling to the large bowlders to keep from being washed into the stream with the receding waters." [1]

In October, 1900, two Geological Survey parties went down the Copper River past Childs Glacier, and Witherspoon made a very good topographic map of it.[2] His map (Fig. 58), as well as photographs taken by A. C. Spencer at about the same time, show Childs Glacier surface and profile, and also its front in Copper River essentially as in 1909.

Since 1900 many photographs of Childs Glacier have been taken by surveyors, geologists, prospectors, tourists, and others, in connection with trips up and down Copper River. Among these may be mentioned the photographs by Webster Brown in August, 1905, the winter photographs of Childs Glacier cliff by J. L. McPherson in January, 1906,

[1] Wilson, H. L., Jr., Copper River Exploring Expedition, 1899, Washington, 1900, p. 50.

[2] Geology and Mineral Resources of a Portion of the Copper River District, House Doc. 546, 56th Congress, 2nd Session, Washington, 1900, Plate II; map republished as Pl. I, Bull. 374, U. S. Geol. Survey, 1909, and as Chitina Quadrangle.

photographs by F. H. Moffit in 1906, the photographs of ice falls and advancing waves by Rex Beach and Fred Stone in 1908, and the photographs by E. A. Hegg, a professional photographer at Cordova. Beach [1] has written a vivid description of the glacier and river.

Observations by the Railway Engineers. Between 1906 and 1910 the Copper River and Northwestern Railway was being surveyed and built, and Mr. E. C. Hawkins, the chief engineer, furnished us with a valuable body of precise data in the form of maps, photographs, and personal observations by himself and his assistants.[2] This information is especially useful (a) because one of the early proposed railway routes crossed the Copper River at the southern end of the Childs Glacier cliff; (b) because the engineers remained a long time at the bridge less than half a mile east of the northern terminus, where the bridge engineer Mr. O'Neel kept precise records of river conditions and meteorological changes from 1907 to 1910, besides taking many photographs; and (c) because Mr. A. O. Johnson, one of the assistant engineers, took an especial interest in the behavior of the glaciers, and saw to it that precise maps were made during the unusually important period in the summer of 1910. To Messrs. Hawkins, O'Neel, and Johnson, and many others, we are under special obligations for turning over to us freely the results of observations between 1906 and 1910.

Observations by the National Geographic Society's Expeditions. Our own observations of Childs Glacier were made in August, 1909,[3] in August and early September, 1910,[4] and in June, 1911.[5] In 1909 we only saw the glacier in passing on the railway, but in 1910 the junior author spent about two weeks in the vicinity of Childs Glacier, making careful studies of general conditions of glaciation and of special phenomena at the time of an unusual advance. Many photographs were taken, and Mr. Lewis made the topographic map reproduced as Map 9 (in pocket). In June, 1911, four days were spent at Childs Glacier by the junior author.

Normal Condition from 1884 to 1905. From 1884 to 1906 Childs Glacier is believed to have terminated in Copper River with a convex front, the amount of ice brought forward from the snowfield being about balanced at the terminus by ablation and by the discharge of icebergs that were carried away by the river. This belief is supported by such photographs, maps, and observations as are available. Allen's description of the glacier in the winter of 1885, with the river about 375 feet wide and the ice so rough that it was impossible to travel upon it, suggests a normal winter condition when lack of melting and of iceberg discharge result in advance of the ice cliff, narrowing the channel to much less than the usual summer width and breaking up the river ice in

[1] Beach, Rex, Everybody's Magazine, Vol. XIX, 1908, pp. 778–9, 784, 785–6, 789; The Iron Trail, New York, 1913, pp. 207–208; Underwood, J. J., Alaska, New York, 1913, pp. 45–49, 58–59; Ellis, Carlyle, Overland Monthly, Vol. LVIII, 1911, pp. 457-466, Technical World Magazine, Vol. XVII, 1912, pp. 3–13; Culmer, H. L. A., Deseret Evening News, Salt Lake City, Utah, Dec. 16, 1911 (illustrated by reproductions of oil paintings of Childs and Miles Glaciers).

[2] Before his death we had the favor of having Mr. Hawkins read and criticise the manuscript of this chapter.

[3] Tarr, R. S. and Martin, Lawrence, Nat. Geog. Mag., Vol. XXI, 1910, pp. 13, 37.

[4] Martin, Lawrence, Nat. Geog. Mag., Vol. XXII, 1911, pp. 541–548; Mastering the Alaskan Glacier Barriers, Scientific American Supplement, Vol. LXXI, 1911, pp. 305–307; Gletscheruntersuchungen längs der Küste von Alaska, Petermanns Geog. Mitteilungen, Jahrgang 1912, Augustheft, pp. 79-81, Tafeln 9, 10; Two Glaciers in Alaska, Bull. Geol. Soc. Amer., Vol. 22, 1910, p. 731.

[5] Martin, Lawrence, in H. F. Reid's Variations of Glaciers for 1911, Jour. Geol., Vol. XXI, 1913, p. 424; Alaskan Glaciers in Relation to Life, Bull. Amer. Geog. Soc., Vol. XLV, 1913, pp. 803, 807–809.

front of the glacier. Witherspoon's map made in 1900 (Fig. 58) shows a normal convex glacier terminus.

Slight Advance in 1905 and 1906. A slight interruption of the balanced condition of the previous years took place in 1905 and 1906, for, (a) Webster Brown's photograph taken in August, 1905, shows a little crevassing of the hitherto stagnant moraine-covered northern margin; (b) Hegg's photograph of July 20, 1906, shows jagged pinnacles of crevassed black ice in this northern portion, at a time when the middle of the glacier was fronted, at least in part, by a gravel bar; and (c) these photographs and the earliest maps by the railway engineers (Fig. 59) show the northern margin projecting eastward a short distance beyond the middle of the front, giving the terminus a concave shape, contrasting decidedly with the convex terminus as mapped by Hayes in 1891 and Witherspoon in 1900.

That the 1905 and 1906 forward movement was slight is indicated by the maintenance of a fairly uniform average river width from 1900 to 1908, although it is possible that the width remained constant while the glacier gradually forced the river eastward. Even if this was the case, maps prove that the displacement was very slight, and of quite a different order from the 1910 advance.

Normal Condition from 1906 to 1909. From 1906 to 1909 the glacier seems to have resumed the balanced condition during which the steady supply of ice brought forward to the river was approximately equal to that taken away by melting and as icebergs, so that, while forward motion took place all the time in the glacier, its terminus neither advanced nor retreated appreciably.

The photographs by Webster Brown, in August, 1905, and by J. L. McPherson, in January, 1906, show that the middle of the front of Childs Glacier formed a nearly straight north-south line, while the northern margin projected eastward very slightly beyond the middle, which was being cut back to a fairly uniform position by the river. This condition was maintained up to 1909. Sometimes the glacier and sometimes a gravel bar (Pl. CL, A) formed the western bank of the river, the condition varying with the time of year (as in 1898), due we believe, to the variations in river level and the position of its channel rather than to advance and retreat of the ice front. Thus a railway survey made early in the summer of 1905 indicates no gravel bar in front of the glacier, but Webster Brown's photograph shows that a narrow gravel bar was exposed in front of parts of the southern half of the ice front on August 16, 1905; and in October, 1905, one of the railway engineers, Mr. Murchison, saw the whole face of Childs Glacier separated from the river by a gravel bar 40 or 50 feet wide. A map by the railway engineers made August 24, 1907, shows a bar of coarse gravel and bowlders along the whole front of the glacier, with a width of 100 to 300 feet and with shallow water near the western bank. The main channel was then near the eastern bank of the river. The horizontal variation in position of the water margin on the eastern bank was 100 to 150 feet with the 24 foot seasonal rise and fall of the river level.

Beginning of Advance in 1909. The width of the river in front of Childs Glacier remained fairly constant from 1900 to 1908, showing variations in 1909, as indicated below.

After applying a correction for the seasonal rise and fall of the river, the decrease in width of the river from 1908 to 1909 shows an advance of the glacier of from 300 to 500 feet sometime during the summer of 1909, as indicated in the following table:

VARIATIONS IN WIDTH OF COPPER RIVER IN FRONT OF CHILDS GLACIER FROM 1900 TO 1909

Date	*Width of River*	*Based on Maps by*
October 9, 1900	About 1600 feet	D. C. Witherspoon
November, 1905	1100–1650 "	Railway Survey
Summer, 1906	About 1600 "	U. S. Geol. Survey
August 24, 1907	1275–1750 "	Railway Survey
Summer, 1908	1000–1800 "	" "
Spring, 1909	625–1125 "	" "
July, 1909	775–1300 "	" "

During this period of advance, in 1909, the ice motion was about four feet or more per day near the moraine-covered northern margin of Childs Glacier. Nearer the glacier margin, just north of the point of observation (a, Fig. 57) the terminus of the glacier had no appreciable motion; at least the edge did not move forward, for small shrubs still grew upon it, as they had done from 1900 to 1909.

The determination of the rate of motion in 1909, recorded above, is the work of A. O. Johnson, one of the railway engineers, who made instrumental observations for two weeks, which he has kindly given us permission to publish.[1] He used a baseline 2983.7 feet long and made transit readings upon a piece of gas pipe 5 feet long driven into the ice, with a flag on top of it. These 1909 observations show the following movement:

RATE OF MOVEMENT IN CHILDS GLACIER IN JULY, 1909

July 15, 8 a.m. to July 16, 3 p.m.	$1\frac{7}{8}$ inches per hour.
" 16, 3 p.m. " " 18, 7 a.m.	2 " " "
" 18, 7 a.m. " " 20, 1 p.m.	$1\frac{5}{8}$ " " "
" 20, 1 p.m. " " 23, 12 noon	$1\frac{3}{4}$ " " "
" 23, 12 noon " " 25, 8 a.m.	$2\frac{3}{16}$ " " "
" 25, 8 a.m. " " 30, 8 a.m.	$2\frac{1}{16}$ " " "

The average of two weeks' observations is, therefore, a little less than 2 inches an hour, or $45\frac{1}{2}$ inches a day. Mr. Johnson says that the rate of movement at this point was slower than that in the middle of the glacier, and we assume that in mid-glacier it may have been as much as 6 feet a day.

The rate of melting of the surface of Childs Glacier during the summer of 1909, may be computed from the fact that the gas pipe holding Mr. Johnson's flag, and which was drilled into a flat surface to a depth of 3 feet, had to be reset every 5 days from July 15 to 30, because the surrounding ice melted to the bottom of the drilled hole. This would give a minimum rate of melting of a little over 7 inches per day.

Mr. Johnson estimated that the amount and rate of discharge of icebergs in 1909 was about twice as great as in 1908, and he states that the iceberg waves began to cut far

[1] Personal communication, Feb. 12, 1910.

more actively into the river bank than they had done in previous years. A river bank about 25 feet high was cut back for 20 feet in one place, though elsewhere the average retreat of the gravel bluff was not more than 3½ to 5 feet during the summer. When the ice front of the advancing glacier was sufficiently undercut by the river the ice slid down into the water with a splash, forming a wave which rushed with great force across the river and up on the opposite bank. In 1908 one such wave deposited a bowlder weighing approximately 200 pounds in the forks of a tree 30 feet back from the edge of the bank. Owing to the decreased width of the river in 1909, however, the waves dashed back over the bank for a much greater distance than in 1908, tearing away earth, bowlders, and trees, and breaking or knocking the bark from trees 60 feet from the river.

That the 1909 rate of motion was slow compared with that a year later is indicated by the fact that the engineers were able to go out on the glacier every few days in July to reset the flags for measuring the rate of motion, while in 1910 this portion of the ice was so severely crevassed that this could not be done. From these facts Mr. Johnson concluded that the glacier was pushing forward and forcing the river eastward in the summer of 1909; and this commencement of advance was announced by us after our visit to Childs Glacier in 1909.[1]

Spasmodic Advance in 1910. The rate of motion in Childs Glacier increased during the winter of 1909–1910 so that part of the margin, hitherto stagnant, attained a forward movement of from two to eight feet a day, while in mid-glacier the rate of movement increased at least five or six times.[2]

The ice which covers Copper River in winter serves as a delicate index of the rate of motion of Childs Glacier during that season. The railway engineers testify that in winter the river ice has always been broken more or less by advance of the glacier, and the description by Allen in 1885 and a photograph by McPherson in January, 1906, shows that there was slight breakage of the river ice at these times. During the winter and early spring of 1909–10, the glacier moved forward rapidly enough to buckle up the ice of the frozen river much more than usual, which is interpreted as proof that the advance of Childs Glacier continued throughout that season.

By June 10, 1910, the ice front in various portions near mid-glacier had moved forward from 275 to 650 feet since the preceding autumn; but the total advance, including that of 1909, was about 1200 feet. Between the measurements of July, 1909, and June 3, 1910, the river had been narrowed down to a width of from 500 to 650 feet, and, if by this constriction the eastern river bank had been cut back, the amount of advance stated above is to be considered a minimum to which it may be necessary to add from 40 to 60 feet. As a result of the constriction resulting from this advance the water rose 21 feet in a few days.

People locally speak of the discharge from the front of Childs Glacier as "sloughing." A "slough" has always raised waves in Copper River (Pl. CXLIX), making it dangerous for a boat to shoot the rapids in front of the glacier, or to line a boat up the opposite bank; but in the spring of 1910 the dangerous conditions were accentuated by the advance of the glacier and the pushing of the river eastward. In that year the waves

[1] Martin, Lawrence, in H. F. Reid's Variations of Glaciers for 1909, Journ. Geol., Vol. XIX, 1911, p. 88; Zeitschrift für Gletscherkunde, Band V, 1911, pp. 200–201.

[2] Martin, Lawrence, Nat. Geog. Mag., Vol. XXII, 1911, pp. 541–548; in H. F. Reid's Variations of Glaciers for 1910, Journ. Geog., Vol. XIX, 1911, pp. 458–459; Zeitschrift für Gletscherkunde, Band VI, 1911, pp. 102–103.

washed up over a bank 5 to 25 feet in height and rushed back 100 to 200 feet into the alder thicket, which is much farther than the waves reached in 1909. Much gravel and sand, stones a foot or two in diameter, and ice blocks (Pl. CL, B) up to ten tons in weight were thrown in among the trees. Alders, 9 to 11 inches in diameter, were stripped of leaves and bark, and bent backward or broken off short, or uprooted, or buried beneath the gravel and bowlders. The river bank, which at some points of measurement was cut back from ½ to 5 feet in 1909, was fairly eaten up by the iceberg waves which crossed the river in 1910, as is proved by measurements at the same points along the bank of the stream facing the glacier, where the recession of the river bank was from 40 to 60 feet.

The amount of retreat of the river bank is shown in the following tables, the 1909 measurements having been made by the railway engineers, while those of 1910 were made by our topographer from points on the same pair of short baselines which Mr. Johnson established the year before. The material of the river bank is very coarse, unconsolidated, glacial gravel and bowlders. The points of observation are on the eastern bank of Copper River, nearly opposite the middle of Childs Glacier.

Measurements from Baselines to Bank of Copper River in 1909 and 1910

Place of Measurement	*July 19, 1909*	*Aug. 1, 1909*	*Oct. 23, 1909*	*Aug. 18, 1910*
Pt. A to river bank	44 feet	41 feet		In river
Pt. B " " "	63 "	58 "		In river
Pt. E " " "	67 "	66 "	65½ feet	16 feet
Pt. F " " "	54 "	49 "	45½ "	–15 feet

Pt. E is the only one of the points of measurements that was left in 1910, the river bank having receded beyond the sites of the other three, though in the case of Pt. F the position was determined exactly by measurement of direction and distance from Pt. E. These measurements show that the increase in iceberg discharge in the river, which was narrowed by the advance of the glacier, resulted in retreat of the river bank as follows:

Amounts of Recession of River Bank Through Cutting by Iceberg Waves in 1909 and 1910

Days Between Observations	*Pt. A*	*Pt. B*	*Pt. E*	*Pt. F*
July 19 to Aug. 1, 1909, 13 days	3 feet	5 feet	1 foot	5 feet
Aug. 1 to Oct. 23, 83 days			½ foot	3½ ft.
Oct. 23 to ——,[1] 1909, ——[1] to Aug. 18, 1910 —days	Over 41 ft.	Over 58 ft.	49½ ft.	60½ ft.

In addition to the retreat at these points there was in 1910 a general retreat of the whole river bank facing the glacier, shown particularly well by the undercutting and

[1] Dates between which the river was closed by ice, so that wave work ceased, though there may have been some erosion of the river bank by the ice blocks which were shoved against the river during the winter.

destruction of a well-beaten path, which formerly followed the edge of the river bank for some distance. Shrubs formerly inside this path were everywhere falling over the edge of the undercut river bank in 1910.

About the middle of August, when the waves were not as powerful as they had been

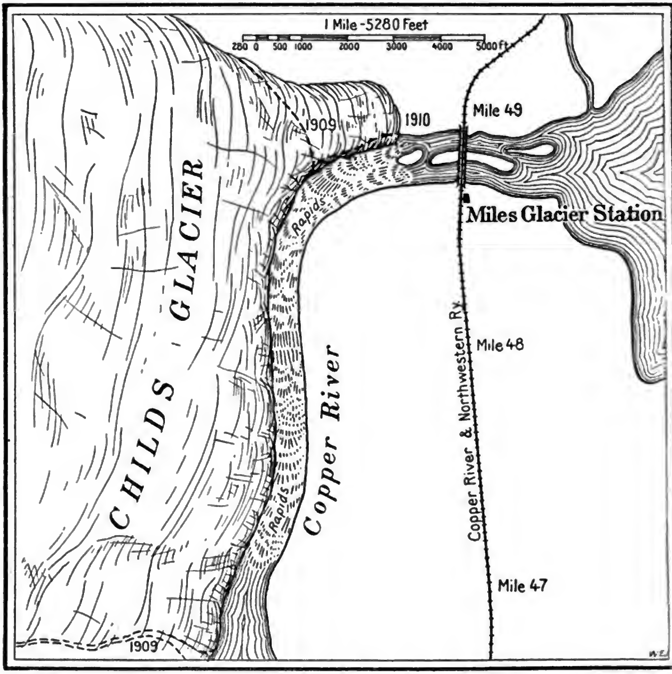


FIG. 59. CHILDS GLACIER IN 1909 AND 1910.
Note the tongue of ice on the north edge, threatening the railroad bridge, in 1910.

earlier in the summer, the largest ice fall, or "slough," which we saw sent a wave across the river with lightning rapidity, causing the water to wash up over the 15 to 20 foot bank near where we stood and to extend back a short distance into the forest. The wave, of course, rose highest directly opposite the ice fall, its height on the bank decreasing both up- and down-stream. Six or seven ring waves in succession splashed far up the river bank, the first one visibly running up-stream, against an eight or nine-mile-an-hour cur-

rent. With this before our eyes it was easy to see how the still larger waves during the earlier part of the summer were able to cut the river bank so easily.

The ice cliff in mid-glacier was about 285 feet high in August, 1910, parts of it having irregular projections with towers and spires of ice, and all of it being severely crevassed. A comparison of photographs of 1909 and 1910 shows that the terminus of the glacier was considerably thickened by the advance. At the time of our visit in August, 1910, the river was so low that some of the ice falls lay where they fell, though many were immediately carried away by the river. At one point at the base of the ice cliff the water was so shallow that large bowlders rose above its surface, and it was evident that undercutting was then much less active than when a deep current swept the glacier face; but it was still effective enough to cause the ice front to retreat while the glacier was flowing strongly forward. We heard many more ice falls in the evening and night than during the day, the rumbling and roaring from the glacier increasing every night after sunset. This may possibly be due to greater contraction of the ice at night, resulting in breaking and scaling off of ice fragments, so that undercutting by the river resulted in great falls. The same increase in noises from the glacier at night was also observed at Columbia Glacier in 1910 and at Childs Glacier in 1911.

The southern border of the glacier was moderately crevassed in August, 1910, the cracks having extended throughout the lateral zone of ablation moraine and to the very edge of the glacier. There was much thickening with the advance (Pl. CLI) and enough lateral spreading to cover the site of a former marginal stream, the glacier terminating close to the edge of an alder-covered bluff.

Near the northern margin of the glacier is an easily-accessible portion of the ice front, which ends upon a nearly-flat plain of till and glacial gravels, overgrown with alder and cottonwood trees 50 to 150 years old (Pl. CLII). Here, as in mid-glacier, the ice advanced slightly between 1900 and 1906, probably in 1905–6, but did not move forward more than a hundred feet or so along this margin. It was nearly without motion from 1906 to 1909, so that small shrubs had begun to grow upon the stagnant margin of the glacier which projected eastward only slightly beyond the main ice cliff. This part of the glacier advanced over 1600 feet between 1909 and June 3rd, 1910, and 204 feet more up to October 5th of that year, when there was a conspicuous east-projecting lobe (Fig. 59). The following table of distances from the glacier margin to the site of the northern end of the railway bridge (Pl. CXLVIII, A) shows the advance of this northern lobe:

Advance of Northern Margin of Childs Glacier Toward Railway Bridge

Date	*Distance from Northern Margin of Glacier to Bridge*	*Nature of Change*
Summer, 1906	3200 feet	Slight advance since 1900
Aug. 24, 1907	3200 feet	No change appreciable
Spring, 1909	3440 feet	Slight retreat
June 3, 1910	1775 feet	Advance 1665 feet
Aug. 17, 1910	1624 feet	Advance 151 feet
Oct. 5, 1910	1571 feet	Advance 53 feet
June 16, 1911	1474 feet	Advance 97 feet

During the summer of 1910 part of this margin was accurately mapped eight times, seven of the surveys being made by the railway engineers, and one by our topographer. The changes are shown in the accompanying table and in Pl. CLIV.

RATES OF ADVANCE OF NORTHERN MARGIN OF CHILDS GLACIER IN 1910 AND 1911

Dates	*Days*	*Advance, in Feet*		*Rate per Day, in Feet*	
		Greatest	*Average*	*Fastest*	*Average*
June 3 to July 29, 1910	56	124	116	2.2	2.07
July 29 to Aug. 6	8	26	23	3.25	2.87
Aug. 6 to Aug. 11	5	41	8	8.2	1.60
Aug. 11 to Aug. 17	6	27	4	4.5	.66
Aug. 17 to Aug. 29	12	42	19	3.5	1.58
Aug. 29 to Sept. 19	21	37	27	1.76	1.28
Sept. 19 to Oct. 5, 1910	17	13	7½	.7	.44
Oct. 5 to June 16, 1911	254	...	97	....	.38

This striking rate of advance, averaging two to eight feet a day, shows variations in velocity with two maxima in the average movements. The fastest movements show a very rapid increase in rate of advance to the maximum August 11, and an unusually rapid decrease again.

It was a rare opportunity to witness the visible forward movement of this margin of Childs Glacier into the forest, as we did during daily visits about the middle of August. A series of lobes developed, though some of them were not persistent (Pl. CLIV), and at the ends of these lobes the day-to-day changes were most pronounced. Enormous ice blocks rolled down the frontal slope, some of them sliding many feet into the forest. Trees were overturned. Turf and grass were ploughed up and carried on the glacier (Pl. CXLV, B), whose ice blocks acted as ploughshares, which ripped up turf and shrubs and carried them along ten or fifteen feet higher than the level at which they grew on the plain. Gravel and turf were heaped up in a discontinuous terminal moraine. Yet one saw and heard little of a spectacular nature while traversing the ice front. It was an irresistible, steady movement; but slow, as the mills of the gods. As impressive as anything was to find tons of ice resting where one stood to take a photograph the day before, or to find some great tree 100 years old prone on the ground with the butt beneath the glacier, where the day before the tree was upright, with the ice just touching it, or with a space of from one to three feet between the glacier and the tree. In places ice blocks, after sliding forward, lay tilted at all conceivable angles on top of the push moraine; but this seemed to be clearly overriding, rather than submarginal accumulation.

Near the river, the ice was thickly covered with ablation moraine and had relatively few crevasses in August, 1909; but photographs taken during the winter of 1909–1910 and early in the spring of 1910, show it severely riven by cracks. In August, 1910, practically all the moraine had slid into the crevasses, though a few small stones still

lay upon the serac tops (Pl. CLV). These were, at the time of our visit, continually sliding down into the cracks, for early in 1910, this margin was so little broken that some of the bridge engineers had no difficulty in walking over it; but by August it was so crevassed that we found it impossible to ascend the precipitous glacier margin or to walk more than a few yards upon the ice. We, therefore, assume that nearly all of the crevassing took place during the spring and early summer of 1910.

Despite the great forward movement, however, the lateral spreading did not result in severe crevassing in all of the northern margin of the glacier, for the extreme northern margin was still thickly covered with ablation moraine (Pl. CLIII) with crevasses extending into the edge of it and not affecting the outer portion, where a few small shrubs still grew upon the ice in August.

There were notable variations of drainage along the northern margin. In 1909, and for some time before, there was a small marginal stream in front of the glacier, which flowed southward to the Copper River at high water, entering the river directly; but at low water stages, the main portion of this marginal stream flowed eastward, behind a gravel bar, nearly to the railway bridge before entering the westward-flowing Copper River. On June 10, 1910, this stream was at the very edge of the glacier, but a little earlier some of the engineers observed that the glacier was 75 to 100 feet west of the stream. The stream did not maintain a marginal position in front of the advancing glacier, however, for on July 29 there was no stream visible. At the time of our visits, in August, a marginal lake had formed, covering part of the northern alluvial fan and extending eastward nearly half a mile into the forest, where the water stood waist-deep among the trees. This lake ended 200 or 300 yards north of Copper River and had an invisible subglacial outlet. Between September 19 and October 5 the outlet, then visible, was extended over 200 feet southward, cutting back the glacier margin some distance, and then continuing in a subglacial course to the Copper River.

The rate of movement in mid-glacier during the summer was not determined, but judging by the measured rate at the northern margin, and by comparison with measured rates at intervals across glaciers in other regions, it may well have been at least six times the 1909 rate, that is 30 or 40 feet a day, or at times even more. But the middle of the glacier did not advance as far during the summer as the northern margin did, and during 16 days in May, when the river was rising, it retreated 200 feet, while between June 3 and August 11 there was a retreat of 415 feet, and between August 11 and 17 of 160 feet. This retreat occurred while the northern margin was advancing most strongly and it seemed to us that it was not due to a cessation of forward motion but wholly to undercutting by the Copper River, which rose six feet between June and August. We accordingly predicted in August that the ice front in the river would readvance during the latter part of September when the river was lower. This prediction proved correct, for up to October 5 there was an advance of 325 feet in mid-glacier plus whatever additional retreat occurred between August 17 and the date when the advance commenced, the river having fallen over nine feet in the meantime. The rate of retreat from August 11 to 17 was at least 27 feet a day.

Since the glacier advanced at least 325 feet in the 49 days between August 17 and October 5 the rate of actual advance in mid-glacier was over 6½ feet a day. As (a) icebergs were discharging all the time, as (b) the advance was surely over 325 feet, and as (c) the forward movement began not on August 17 but probably after the

middle of September, when the river was much lower, the rate of advance undoubtedly exceeded 6½ feet a day. If the advance commenced on September 19, when the river level lowered, the forward movement of more than 325 feet in the 16 days before October 5 was at the rate of over 20 feet a day. Taking the minimum of 6½ feet a day, however, the following computation may be made. If the middle of the front of the glacier moved forward at the rate of at least 6½ feet a day when the northern margin was advancing less than half a foot a day (September 19 to October 5, see table, p. 406) then the middle of the glacier may have been moving over 40 feet a day between July 29 and August 6 when the northern margin had an average velocity of nearly 3 feet a day. Repeating the same calculation on the basis of movement at the rate of 20 feet a day in mid-glacier up to October 5, it appears that the ice in the middle of Childs Glacier may have been flowing at the rate of at least 130 feet a day between July 29 and August 6. These figures are, of course, mere estimates, but they indicate clearly the order of magnitude of the rates of glacier motion during a spasmodic advance. The ice was too much crevassed to permit the setting up of flags and there were no pinnacles of ice that could be identified from day to day, so that we were unable to make precise measurements of the rate of movement. It seems clear to us, however, that a rate of movement of not less than 30 or 40 feet a day occurred in mid-summer of 1910, in the central part of the glacier; and the rate may well have been several times that amount.

The oscillations of the ice front in mid-glacier during the summer of 1910 are shown in the following table:

Variations in Width of Copper River in Front of Childs Glacier with Oscillations of Ice Front in 1909–1910

Date	*Width of River*	*Advance or Retreat*	*Based on Maps by*
July 1909	775–1300 feet	Beginning to advance	Railway Engineers
June 3, 1910	500–650 feet	Advance 275–650 feet	" "
Aug. 11, 1910	1065 feet	Retreat 415 feet	" "
Aug. 17, 1910	625–1225 feet	Retreat 160 feet	Nat. Geog. Society
Unknown date	Not measured	Stationary	Not surveyed
Oct. 5, 1910	600–900 feet	Advance 325 [1] feet	Railway Engineers

What the glacier might do before the spring of 1911 was of great interest. It might continue to advance; or, as the diminishing rate of advance on the northern margin after August 11 suggested, it was possible that the strongest advance was over. If advance continued, would the glacier move up to and destroy the railway bridge (Pl. CXLVIII, A), which was only 1571 feet distant from the northern margin (Fig. 59), or would it stop before advancing that far? The bridge cost $1,500,000, and is the key to the new $20,000,000 railway to the copper mines. It is absolutely certain that no corps of engineers living could save the bridge and railway if the glacier should advance that far. The railway was less than a mile from the middle of the glacier, which might easily have advanced this distance between May and October,

[1] Plus unmeasured retreat after August 17.

1910, when it was moving at the rate of 30 to 40 feet a day, or more. If the glacier terminus had not been swept by the Copper River, the rapidly-moving middle of the glacier certainly would have advanced this distance. The slowly-moving northern margin did advance 1869 feet, at a rate which increased from practically zero in 1909 till it reached 2 to 8 feet a day in 1910. Advance here was possible because this part of the glacier ends on the land where the river could not cut it back.

At the end of the summer of 1910 this northern portion of the glacier was only 1571 feet from the railway bridge. If the river had not had its normal summer rise, or had the advance occurred in the winter when the water was low and the river weak, the middle of the glacier would surely have advanced a good part, if not all, of the distance to the railway. As it was the ice front did move forward more than 600 feet before the rise came. It is evident, therefore, that the preservation of the railway and bridge during the summer of 1910 was due to the river, the very thing which necessitates the bridge. The relationships of the variations of the glacier to stages of river level are as follows:

Oscillations of Middle of Childs Glacier and of Copper River Level in 1910

Date	*Variations of Glacier*	*Stage of River*
1909 to June 3, 1910	Advance 275–650 feet	Fall [1] 12 feet, May 6 to June 3
June 3 to Aug. 11	Retreat 415 feet	Rise 6 feet
Aug. 11 to Aug. 17	Retreat 160 feet	Level about stationary
Aug. 17 to unknown date	Retreat, continued	Rise $1\frac{7}{10}$ feet
Unknown date to Oct. 5, 1910	Advance 325 feet [2]	Fall 9 feet

Resumption of Normal Conditions in 1911. Upon the return of the junior author to Childs Glacier in June, 1911, it was found that the slowing down of the advance of the previous year (see table, p. 406) had been continued. The northern margin of the glacier had advanced only 97 feet toward the railway bridge (Pl. CLIII), and the rate of motion had been reduced from 44-100 of a foot a day in September–October, 1910, to 38-100 of a foot a day in June, 1911. Conditions were not yet quite normal, however, for under usual conditions this portion of the glacier terminus has little motion, and the amount of ice brought forward is so nearly balanced by melting, that the ice edge either remains stationary or retreats. In 1911, however, the northern margin was still advancing slowly, but appreciably, into the forest, a tape line measurement proving an advance of approximately 1 foot in the 3 days between June 16 and 19. This portion of the ice front was almost as severely crevassed as in the previous summer. There was much less terminal moraine and in most places there was none, though at one point the moraine was 5 or 6 feet high. On the river bank there was a low, partly-submarginal moraine, made up of rounded bowlders and including some dead alders. The stream draining the marginal lake flowed through the forest parallel to the ice front, having abandoned its subglacial channel since the previous autumn.

That the rate of motion decreased also in mid-glacier is made plain by the fact that

[1] Low water throughout most of fall and winter, with rise from March 16 to May 6.

[2] Plus unmeasured retreat after August 17.

the ice front did not advance completely across the river channel during the winter; but that it did advance all winter is proved by the breaking of the river ice, observed by the watchman at the bridge. The channel was 600 to 900 feet wide in October, 1910, and approximately 400 or 500 feet wide in June, 1911. During most of the intervening months the river was so low that a gravel bar separated the ice cliff from the river. Therefore, there was practically no loss of ice through undercutting and discharge of icebergs, and during the winter months there would be little loss from ablation. Yet the glacier did not advance completely across the river channel, as it would have done if it had been moving as much as $6\frac{1}{2}$ feet a day, as was the case in the previous autumn. In fact, the rate during the winter was less than 2 feet a day.

Exactly how much the ice front did advance during the winter is not known. It was at least 200 to 400 feet and probably not much more. The river did not rise high enough to cover the gravel bar until sometime between June 6 and 9, 1911. Then the undercutting of the ice cliff by the river resulted in the beginning of rapid iceberg discharge, spoken of by people in Alaska as the "working" of the glacier. The ice cliff accordingly began to retreat and the width of river observed on June 17 had undoubtedly been increased by some retreat during the previous week. Even with this allowance it is evident that there had been a notable diminution of rate of advance during the winter of 1910–11. Judging by the incomplete data that we have, which at least give the approximate relative magnitudes of the motion, the middle of Childs Glacier was moving at the rate of approximately 2 or 3 feet a day from 1906 to 1908, increasing to 5 or 6 feet a day in 1909, to at least 30 or 40 feet a day in August, 1910, and at the height of the advance perhaps much more. The rate also decreased rapidly, being between 6 and 20 or more feet a day in October, 1910, and less than 2 feet a day in June, 1911.

Signs of the preceding summer's activity were still visible on the river bank opposite the middle of the glacier, and, owing to a rapid rise of the river during the period of observation, the junior author was again able to witness the phenomenon of frequent and active iceberg discharge. The ice cliff was still so near that the waves occasionally rushed across the river and rose over the gravel bluff. Ice blocks of many tons in weight were left by the waves and stones of great size were hurled back among the trees, while some of the smaller shrubs were being uprooted by the waves. Several times these waves were seen to run up-stream from $\frac{3}{4}$ to $\frac{7}{8}$ of a mile against the river current. Hundreds of salmon were washed out upon the bank of the river near the glacier. At the base of the ice cliff the water sometimes splashed 200 feet into the air, and air vibrations in connection with icebergs discharge shook a frame building $\frac{3}{4}$ of a mile from the glacier vigorously, so that the writer was awakened from sleep several times during the night. As in 1910, iceberg discharge was more frequent in night than in the daytime. Yet, on the whole, the phenomena observed in 1911 were feeble compared with those of the previous year when the maximum advance was in progress.

The diminution of movement was also shown by the inactivity of the southern third of the glacier front. Some lateral spreading of the southern margin of the glacier occurred between August, 1910, and June, 1911, but that it had ceased for some time was indicated by the presence of a narrow strip of thin ablation moraine on the ice edge and by the fact that for some distance from the margin the crevasses in the glacier had nearly healed by ablation.

Altogether the 1911 conditions indicated a rapid return to the normal state which prevailed before the advance of 1910.

It would have been scientifically interesting, though unfortunate for the railway, to have seen the advance continue during the low water stage from October, 1910, to June, 1911, with a gradual forcing of the river eastward, the cutting back of the eastern river bank in the unconsolidated gravels, the interference with stream flow and iceberg discharge from Miles Glacier, and many accessory phenomena. Fortunately the glacier advance came at the time of high water, and was nearly over by the beginning of the low water stage in October. The advance would have easily reached the railway if it had occurred in early spring, late fall, or winter. It seems probable, however, that the river can always be depended upon to protect the bridge and railway, if future advances come at high water, when undercutting of the glacier seems competent to cause enough iceberg discharge to keep back the advancing ice front; unless, of course, there occurs a period of prolonged or far more active advance. It is unfortunate that there is no similar protective relationship of the river to the Allen, Grinnell, Heney, and Kennicott Glaciers, whose outer portions are also traversed by this railway and which may sometime have advances similar to that of Childs Glacier in 1910.

Cause of the 1910 Advance. The advance of Childs Glacier in 1910 is of interest in connection with the spasmodic advances or glacier floods of the Yakutat Bay glaciers since 1899. The recent history of the ice tongue may be summarized as follows: Omitting the imperfectly-known history of Childs Glacier at the time of the visits by the Russians, we know that it was engaged in what seems to be normal flowage from 1884 to 1909, interrupted by a slight increase in rate of movement in 1905–6. In 1909 this normal movement was interrupted by a marked increase in rate of flow but this did not assume great proportions until the summer of 1910. During this summer the advance was phenomenal, being similar to what we infer to have taken place in the Yakutat Bay glaciers during their brief spasmodic advances, though with a longer total period of advance, in this respect resembling the conditions in the Vernagt-Ferner of the Tyrol. The Childs Glacier began its unusual activity in 1909, was many times more active in 1910, slowed down by October, 1910, but continued a slight advance during the winter of 1910–11. In June, 1911, the conditions seemed to be nearly normal again, though there was still some advance.[1] This shows clearly that a great wave of motion surged down through Childs Glacier, beginning to make itself felt in 1909, causing greatest activity in 1910, and becoming greatly reduced in about a year from the time it commenced.

Mr. E. C. Hawkins was told by a reliable prospector, who witnessed the happening, that during an unusually warm, rainy period in January or February, 1910, a detached hanging glacier fell from the walls of the Childs Glacier valley about 10 miles west of Copper River. A stupendous quantity of ice was avalanched out upon the glacier surface. This of itself could not have caused the advance of 1910, which indeed had already commenced before the avalanche. It illustrates, in a small way, however, what may take place on a much larger scale when a severe earthquake shakes a mountain region like that near Childs Glacier, which has many detached glaciers hanging on steepened slopes and much snow in unstable equilibrium.

[1] Caleb Corser of Cordova observed in 1912 that Childs Glacier had advanced 60 feet from July 15 to September 30. It is presumed that this was measured at the northern margin (see table, p. 405).

It is known definitely that this portion of the Chugach Mountains had severe avalanches during the Yakutat Bay earthquakes of September, 1899, for the shocks were felt with much severity on the Copper River delta southwest of Childs Glacier, as well as to the southeast; and many avalanches were heard in the Chugach Mountains to the northwest.[1] In all probability avalanches were also caused by the smaller Chugach earthquake of October 9, 1900, which was felt with much severity on the Copper River delta,[2] and indeed by one party between Childs and Miles Glaciers. The rapidity of movement of this wave of advance in 1910 and its rapid diminution strongly suggests the earthquake avalanche origin.

The climatic situation regarding Childs Glacier may be summarized as follows: At Fort Liscum near Valdez, 60 miles northwest of Childs Glacier there was increased snowfall in 1902–1903 and in 1907–1908 (see table, p. 316). At Orca, 33 miles southwest of Childs Glacier, the snowfall[3] was much heavier than usual in 1902–1903 and from 1906 to 1908. For the snow year 1907–1908 the railway engineers' records at Miles Glacier station show a snowfall of 433 inches and for 1908–1909 of 410 inches. In the latter year the snowfall at Flag Point, 20 miles south of Childs Glacier was 355 inches. During the first half of the snow-year 1909–1910 the snowfall at Miles Glacier station was 102½ inches, or only half as much as in the corresponding months of the two previous years. This suggests that the snowfall in 1907–1908 was exceptionally great. The slow advances of Miles and Grinnell Glaciers in 1910, of Heney Glacier in 1911, and of Allen Glacier in 1912, described on subsequent pages of this book, suggest that the advances of all five of these glaciers may be due to increased snowfall in 1902–1903 or 1906–1908, but the junior author does not feel that a sufficient body of fact is as yet available for choosing between the climatic and earthquake hypotheses.

The Terminal Moraine of Childs Glacier. There is unmistakable evidence that Childs Glacier formerly spread out in a great bulb, similar to the present bulbs of Allen, Sheridan, and Miles Glaciers. The proof of this is found in the remnants of the great terminal moraine of Childs Glacier. There are five of these residual masses of glacial drift around the periphery of the former bulb (Pl. CLXV) and m, m (Pl. CLVI). The moraine is an ancient one, everywhere clothed with dense vegetation, chiefly alder and cottonwood.

The first of these is the 600 foot hill on the western side of the Copper River valley, immediately north of Childs Glacier. It is a curved ridge extending eastward a mile and a quarter from the valley wall. Its crest everywhere rises 400 feet above sea level, or 275 feet above the adjacent valley bottom, and three knobs upon it reach elevations 100 and 200 feet higher. It connects with a 500 foot terrace on the northern slope of the valley of Childs Glacier. This terrace slopes rapidly westward to an elevation of 1200 feet. The moraine is ⅜ to ½ a mile wide and, in places, its surface has pronounced minor irregularities. Its slopes are very steep.

This moraine remnant was at first thought to be a rock hill, especially as it rises much higher than any other moraine accumulation in the Copper River valley. In

[1] Martin, Lawrence, Alaskan Earthquakes of 1899, Bull. Geol. Soc. Amer., Vol. XXI, 1910, pp. 351–352, 356, 364–365, 373, 374; Tarr, R. S. and Martin, Lawrence, Professional Paper 69, U. S. Geol. Survey, 1912, pp. 49–50.

[2] Martin, Lawrence, op. cit., p. 401.

[3] Data assembled by Miss Maude Reid from Weather Bureau records. See unpublished thesis, University of Wisconsin, 1913.

its position, however, it is out of harmony with any other uneroded rock spurs along the Copper River, though perfectly normal as a morainic accumulation of Childs Glacier. Repeated traverses of its slopes have revealed no rock ledges, and no rock in place is found in the existing exposures. The first of these is at the extreme eastern end, where the grading for the Copper River and Northwestern Railway cut into the end of the moraine for ¾ of a mile. Here all of the material was till, with large bowlders imbedded in fine clay. The other exposure is at the western end of the moraine, where Childs Glacier has cut into the edge of the moraine terrace, revealing no rock, but till and stratified gravel. These relationships prove that this is a morainic accumulation and not a rock hill, and it is also clear that this is a moraine built by Childs Glacier and not by the former trunk glacier of the Copper River canyon.

There is a gap of a mile and a half between this large northern ridge and the next two remnants of the terminal moraine of Childs Glacier. They are each small knobs, lying obscurely in the forest between the railway and Miles Glacier. The northernmost is a small hillock, rising to a height of 302 feet, or over 100 feet above the adjacent outwash plain. It lies just in the edge of a more-recent, low, terminal moraine of Miles Glacier (Pl. CLXV) and is made up partly of rounded gravels. The other small residual of the terminal moraine of Childs Glacier is a long, low ridge which probably marks the easternmost portion of the former bulb.

The fourth hilly area which represents part of the terminal moraine of Childs Glacier is fairly large, but does not rise very high above the outwash plain. It extends from the railway to the Copper River, beginning with a single, steep-sided ridge which crosses the railway west of Mile Post 47. Close to the railway, it is a narrow, sinuous ridge, esker-like in places, but generally with knobs and kettles, and, at one locality, divided into parallel ridges which enclose the basin of a large pond. Near the railway it rises 25 to 40 feet above the adjacent plain of outwash gravels, which is perfectly smooth except where trenched by parallel, southwest-trending channels of the former streams from Miles Glacier. This moraine broadens and becomes more irregular as it extends southwestward and, near the Copper River, rises to an elevation of 100 to 200 feet.

The fifth and southernmost remnant of this terminal moraine lies south of Childs Glacier and west of Copper River, which has cut a steep cliff in it. The moraine is 210 feet high and extends close up to the present margin of Childs Glacier (Pl. CLXV).

The expanded bulb of Childs Glacier, which built this terminal moraine, was at least 4½ miles wide and projected 1½ miles farther east in the Copper River valley than the present ice front. The time of its maximum independent expansion was later than the stage of a trunk glacier in Copper River valley, and must have also been a time of diminution of the present bulb of Miles Glacier, which overlaps it in part. The Childs Glacier evidently retained this advanced position for an exceedingly long time in order to produce these large morainic accumulations. The Copper River necessarily flowed then on the eastern side of its valley. The latest shifting of the Copper River toward the western side of the valley, doubtless at a time of expansion of the Miles Glacier bulb, resulted in the cutting away of portions of this terminal moraine, as well as in the destruction of the ice bulb of Childs Glacier. Subsequently the outwash gravels were deposited between the moraine remnants by streams from Miles Glacier, and the Copper River was forced over to its present course, where it undercuts the front of Childs Glacier and prevents its terminus from expanding into a symmetrical bulb.

CHAPTER XXI

MILES AND GRINNELL GLACIERS

Location and Relationships

The Miles and Grinnell Glaciers are located just north of Childs Glacier, 25 or 30 miles above the mouth of Copper River (Pl. XCIII), in the lower part of the canyon by which the river crosses the Chugach Mountains. At this point the Copper River valley is from 2½ to 4 miles wide at the bottom and its walls rise abruptly to elevations of from 3000 to 6000 feet. Farther south the valley walls flare apart, the width being 13 miles near the sea; and the glaciers discussed in this chapter are at the head of the V-shaped indentation with which the Copper River canyon ends.

Here four glaciers enter the valley near by at right angles to its course (Map 9, in pocket), Miles Glacier coming from the east and Childs, Grinnell, and Allen Glaciers from the west. Their termini are from 125 to 200 feet above sea level. Miles and Allen Glaciers extend completely across the main valley, forcing the Copper River first against one mountain wall, then the other, so that it writhes between the glaciers in a sinuous course. Miles and Allen Glaciers, like Childs Glacier, each act as a dam to the river, causing rapids opposite the end of each of the three glaciers, and above these, such relatively slack water as to give rise in each case to a lake-like expanse. The expansion above Childs Glacier is a broad lake, into which Miles Glacier discharges icebergs, but the other expansions are mere enlargements of the width of the river. There is no name for the rapids opposite the end of Childs Glacier, but the rapids opposite the termini of Miles and Allen Glaciers are called Abercrombie Rapids and Baird Canyon Rapids respectively.

Copper River is fed chiefly by glacial streams, receiving the drainage of some of the glaciers on the southern side of the Alaska Range, of a large proportion of the glacier system of the Wrangell Mountains, and of the glaciers along 200 miles of the northern slopes of the Chugach Mountains. To this already great supply is added much water from Allen, Miles, and Childs Glaciers.

After the railway crosses the river by the high steel bridge between Miles and Childs Glaciers it passes over the tip of Grinnell Glacier, and close to the northern part of Miles Glacier near Abercrombie Rapids.

Miles Glacier

General Description. Miles Glacier (Pl. XCIII) flows westward into the Copper River valley from an unexplored source in the Chugach Mountains. Only the lower 15 miles of the glacier have been mapped (Map 9), but much more than this is visible from the railway, and it is probably 40 or 50 miles long. It rises in snowfields, north of Bering and Martin River Glaciers, where Mt. Hawkins[1] and several other peaks rise

[1] Named in 1910 for the late E. C. Hawkins, the engineer who built the Copper River and Northwestern Railway.

to heights of 8000 to 10,000 feet near the head of the Miles Glacier. Within the mountain valley the glacier has a width of two to two and a half miles and it receives at least three tributaries in its lower course. There are one or two rather weak medial moraines, but there are broad lateral moraines; that on the north having a width of an eighth of a mile, while the southern lateral is half a mile to a mile in width. Throughout its visible course, in this mountain valley, this ice stream is severely crevassed.

Where it emerges into the Copper River valley, the Miles Glacier spreads out in a great bulb, the width increasing from 2½ miles in the mouth of the valley to 6½ miles, measured from the northern to the southern edge. This bulb has two quite different portions, a precipitous ice wall on the south, and a moraine-covered piedmont area in

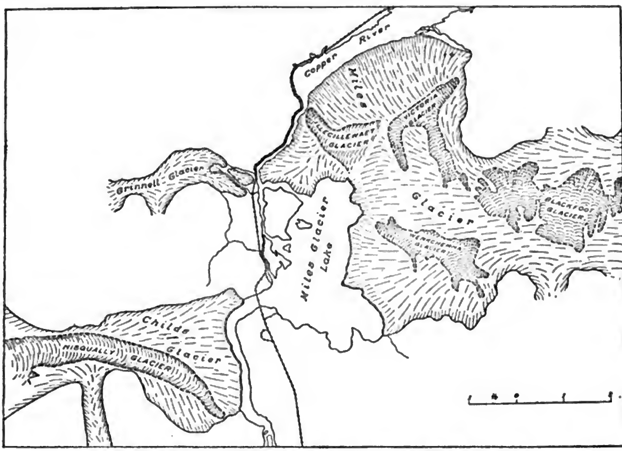


Fig. 60. Map to Illustrate Size of Childs and Miles Glaciers.

the northern portion. The southern half protrudes only slightly from the mountain valley, ending in a vertical, white, ice cliff (Pl. CLVIII, B) which discharges icebergs into the lake-like expanse of Copper River caused by the Childs Glacier constriction just below. This cliff is about three and a half miles long and at least 200 feet high. On the southern border of this clear ice terminus is a detached strip of ice completely mantled by morainic débris, making it impossible to judge exactly where the ice ends. This is opposite the terminus of the broad strip of lateral moraine, barren of vegetation, which extends up the southern border of the glacier.

The northern half of the Miles Glacier terminus, into which the ice cliff grades, projects westward a distance of three and a half miles, forcing the Copper River clear across the valley and into a marginal channel on the western side, where the constriction causes Abercrombie Rapids. This part of Miles Glacier looks so little like a glacier

that it has not been represented as such on some of the maps; for it is stagnant, moraine-veneered, and vegetation-covered, ice showing only here and there where the moraine is slumping (Pl. CLVII) as the unstable foundation slowly melts. Quite in contrast to the ice cliff portion, there is no discharge of icebergs from any part of this moraine-covered margin. It is nearly twelve and a half miles around the periphery of the Miles Glacier front, passing along the ice cliff, then, in a great curve, around the margin of the moraine-covered bulb.

As seen from the railway at Miles Glacier station, the Miles Glacier is less impressive than the Childs, which is nearer. The real difference in magnitude may be appreciated from Fig. 60, which shows the Miles and Childs Glaciers with sketches of several of the larger glaciers of the United States and Canada, on the same scale, superimposed upon them.

Observations by the Russians. Several explorers have quoted the Copper River natives as saying that in early times, before the Russian explorations, Miles and Childs Glaciers were united, with the Copper River flowing underneath. While not at all improbable, there are no known facts to verify this.

As in the case of Childs Glacier the date of the Russian observations quoted by Grewingk in 1850 is not known; nor is it known whether these are Grewingk's own observations or those of some one else. The Russians who are known to have gone up Copper River are Nagaief in 1783,[1] and several others up to the time of Serebrannikoff in 1847.[2] The observer to whom Grewingk refers may possibly have been a sailor like Nagaief rather than a trader like Serebrannikoff and some of the others, for he gives the thickness of the ice in fathoms, not in feet.

Grewingk[3] stated in 1850 that "the ravines filled with ice twenty fathoms thick are a mile wide at the river, and in some places the ice is covered with earth in which grow mosses, berry bushes, and the alder. Frequently we can see the icebergs[4] covered with soil and a growth of green bushes and ripe berries.

"Rapids have been formed above the canyon where the current of the water has cut through the glaciers . . . and beyond these rapids no more ice is found, and the sea breezes and the fogs do not reach."

This seems to be a description of the stagnant, northern part of Miles Glacier at Abercrombie Rapids and perhaps of Allen Glacier, above which there are no other large glaciers that the Russian explorer would have seen from the river.

Observations by Army Officers and Government Geologists. Abercrombie spent July and August, 1884, near Miles Glacier, making a map, taking a number of photographs, afterwards reproduced in Allen's report of the army expedition of 1885, and recording some facts concerning the glacier.[5] Its middle portion was severely crevassed, for in

[1] Bancroft, H. H., History of Alaska, Vol. XXXIII, San Francisco, 1886, p. 187.

[2] Journal of the Russian Geographical Society, St. Petersburg, 1849, translated by S. N. Buynitzi, and quoted by H. T. Allen, in Narratives of Explorations in Alaska, Senate Rept. 1023, 56th Congress, 1st Session, Washington, 1900, pp. 412–413; Dall, W. H., Alaska and Its Resources, Boston, 1870, p. 343.

[3] Grewingk, C., Beitrag zur Kenntniss der Orographischen und Geognostischen Beschaffenheit der Nordwest-Küste Amerikas mit den Anliegenden Inseln, Verhandl. Russ. Kaiserl. Mineral. Gesell. zu St. Petersburg, 1848 und 1849, pp. 34–5.

[4] Probably meaning the glacier rather than icebergs.

[5] Abercrombie, W. R., Supplementary Expedition into the Copper River Valley, Alaska, 1884, Narratives of Explorations in Alaska, Washington, 1900, pp. 386, 389–390.

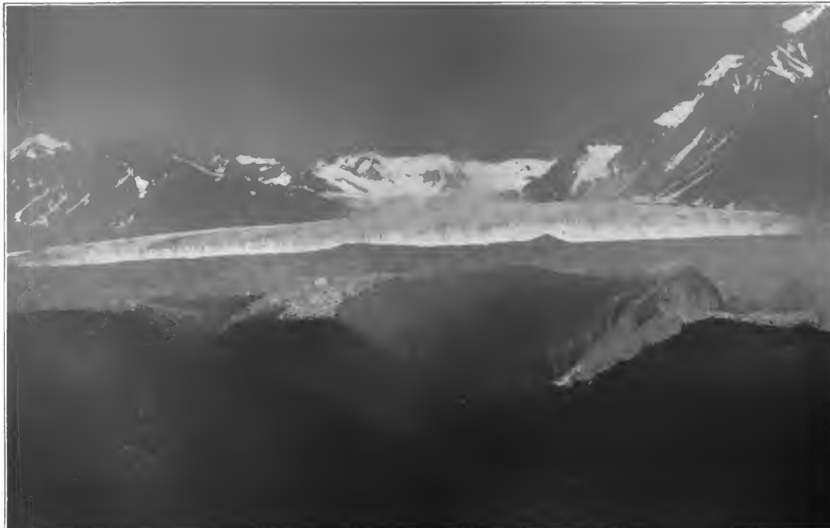
A. Terminal Cascade of Chenega Glacier
Compared in height with Washington Monument, which is 550 feet high. Photograph, August 9, 1910, from Photo Station C.

B. The Advancing Terminus of Childs Glacier
ear the northern edge in 1910, when this portion was advancing into the forest at the rate of about 3½ feet a day. Photograph, August, 1910.

PLATE CXLVI

MAP OF LOWER PORT FIDALGO (BASED ON SOUNDINGS BY U. S. COAST AND GEODETIC SURVEY)
showing the trough produced by glacial erosion below sea level, and the U-shaped, submerged hanging valleys. Submerged contour

Plate CXLVII



Childs Glacier

Showing snowfields, valley glacier and terminal bulb. Photograph, August 19, 1910, from Station G^1 near Miles Glacier. Foreground, ice of Miles Glacier.

A. THE $1,500,000 RAILWAY BRIDGE
Which was menaced by the advance of Childs Glacier in 1910. Photograph from Photo Station H, August, 1910.

B. PART OF FRONT OF CHILDS GLACIER IN COPPER RIVER
Showing comparison of ice cliff, which varies from 250 to 300 feet in height, with the Capitol, in Washington, which is 287½ feet high.

THE CLIFF OF CHILDS GLACIER

Compared in height with the Claus Spreckles Building, San Francisco. The picture shows, on the extreme left, how the icebergs fall from the glacier when the Copper River undercuts the ice cliff, giving rise to great waves. Photograph by E. A. Hegg.

A. LOW WATER STAGE OF COPPER RIVER
With infrequent ice falls, lying on a gravel bar at the base of Childs Glacier cliff.

B. ICEBERGS FROM CHILDS GLACIER
Thrown up among the trees by waves. Photograph by E. A. Hegg.

Childs Glacier
Showing thickening with advance of southern margin from October 15, 1909, to August 17, 1910. Photograph by E. A. Hegg, from Photo Station I, 1909.

PLATE CLII

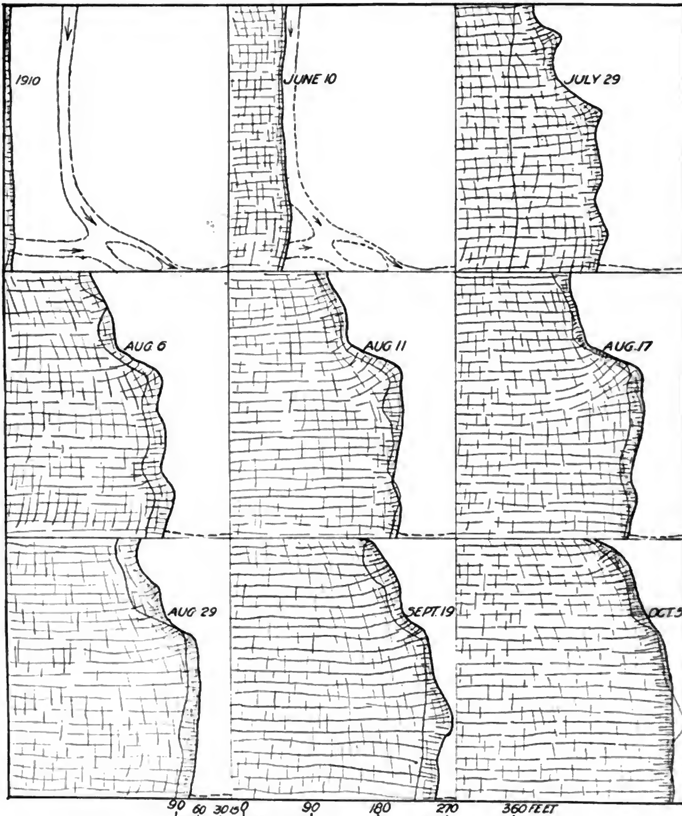
THE NORTHERN MARGIN OF CHILDS GLACIER

In 1909 before the advance, with the 1910 position in ink. Photograph from Photo Station B, June 20, 1909.

PLATE CLIII

THE NORTHERN MARGIN OF CHILDS GLACIER IN 1910

Shows amount of advance and thickening in fourteen months. The advance was nearly 1700 feet. Right hand dot shows additional advance up to June 16, 1911. Mt. O'Neel in background. Photograph from Photo Station B, August 21, 1910.



NINE MAPS OF THE SAME PORTION OF THE ADVANCING FRONT OF CHILDS GLACIER IN 1910
After surveys by railway engineers and the National Geographic Society. On each map the last previous ice front is shown by a light line.

Northern Margin of Childs Glacier
Advancing into mature forest in 1910. Photograph from Photo Station D, on railway bridge, August, 1910.

PLATE CLVI

THE TERMINAL MORAINE OF CHILDS GLACIER

The white cliff of the glacier (g, g) shows on the extreme right. The remnants of the crescentic moraine (m, m, m) consist of (1) the great hill in the foreground, (2) the light-colored hill near the lake, (3) dark-colored hills rising through the smooth valley train of outwash gravels and (4) a light-colored hill, on the right, south of the glacier. This outwash was partly deposited by Copper River, whose present [illegible] of the former [illegible] of Childs Glacier, and partly by streams from Miles

THE MILES GLACIER AT ABERCROMBIE RAPIDS

The whole of the hill across the river is underlain by ice which shows here in three patches. It is overlain by ablation moraine bearing vegetation. The huge bowlders on the other side of the rapids are a part of a heavy terminal moraine. Photograph, from the railway, by E. A. Hegg in 1909.

A. MILES GLACIER, ON LEFT
Holding Copper River in marginal channel of Abercrombie Rapids. Steep mountain wall on right. Photograph, 1909.

B. A PORTION OF THE FRONT OF MILES GLACIER
Mt. Hawkins and Van Cleve Glacier in the background. Photograph, 1910, from Station C, on the railway bridge

The Valley Portion of Miles Glacier and the Terminal Ice Cliff from Station E, August 15, 1910

Plate CLX

The Copper River

In its marginal channel with the vegetation-covered bulb of Miles Glacier on the right. August 10, 1910. From Photo Station N, the site occupied by Abercrombie in 1884.

August they tried to cross it from the south but were turned back. Its northern portion extended clear across the main valley up to the west wall, while the southern part had retreated to form a small part of the present lake. Abercrombie vividly describes the rapids and says that the river was only 150 yards wide, running 10 or 15 miles an hour in midstream. Just below the rapids the river split into four large channels, down which the water rushed "at a terrific rate of speed into a large basin formed by this monster glacier in days gone by." Of Abercrombie's two 1884 photographs, reproduced in Allen's report,[1] one shows the rapids and the ice cliff of the southern portion somewhat as they were in 1910, though the ice cliff was further out; and it also shows almost the same distribution of lateral moraine on the northern side of the glacier as in 1910. The other shows not only that the northern portion of Miles Glacier extended across the valley more than 27 years ago but that it was even then stagnant, and moraine-covered, with vegetation growing upon it, and in process of being undermined by slumping, as in 1909 and 1910.

When Allen[2] went up Copper River in April, 1885, he observed that west of Miles Glacier the river bed was 800 yards wide in contrast with its width of only 125 yards east of Childs Glacier and of 50 yards in Abercrombie Canyon.

Allen[3] recognized that the vegetation-covered hills between Childs Glacier and the southern margin of Miles Glacier, were morainic and associated with a former expansion of these glaciers, indeed stating that the drift grades indistinguishably into the southern edge of Miles Glacier, as we observed it to do in 1909.

When Hayes descended Copper River in August, 1891, he made a sketch map showing the specific conditions at Miles Glacier, which he also described as follows:[4]

"A couple of days brought us down to Miles Glacier, where the river tumbles over a dam of huge moraine bowlders. It is necessary to make a portage here sometimes across both moraine and glacier. Crossing about two miles of moraine covered with a dense alder thicket, we came out upon a high ridge of freshly deposited bowlders. Immediately in front was a broad expansion of the river in front of the glacier, which formed an ice cliff along one side nearly four hundred feet in height. Bergs were almost constantly falling, with reports like thunder, dashing the spray high above the top of the cliff. The current of the river sets across the lake toward the front of the glacier, and where it meets the swell produced by a falling mass of ice the water is thrown into enormous breakers which, with the grinding icebergs, would swamp a boat instantly."

"Miles Glacier is quite comparable in size with those of the St. Elias region and is formed under essentially the same climatic conditions. It is evidently retreating at present, and the river spreads out in a lake-like expansion along its front in a part of the glacial channel from which the ice has receded. This expansion of the river is about a mile in width and one side is formed by the glacier front, a cliff of ice 350 feet above the water and over five miles in length. Although the ice no longer reaches entirely across the valley, there remains a heavy lateral moraine, indicating its former position

[1] Expedition to the Copper, Tanana, and Koyukuk Rivers, 1885, Washington, 1887, Plates 5 and 6; Narratives of Explorations in Alaska, Washington, 1900, opposite pp. 426 and 427.

[2] Allen, H. T., Expedition to Copper, Tanana, and Koyukuk Rivers, Washington, 1887, p. 42, and Map 2; also in Narratives of Explorations in Alaska, Washington, 1900, pp. 424–5, 486–7.

[3] Allen, H. T., Copper River, Alaska, Glacial Action, Science, Vol. VIII, 1886, pp. 145–146.

[4] Hayes, C. W., Exploration Through the Yukon District, Nat. Geog. Mag., Vol. IV, 1892, pp. 126, 154.

and damming back the river as already described. The fact that the river has cut only part way through the moraine indicated a very recent recession of the glacier."

Although no reference is made to ice beneath the moraine, there can be no doubt that it was there in 1891, as it was in 1909.

Abercrombie revisited Miles Glacier and Abercrombie Rapids in October, 1898, fourteen years after his first visit, and states that "the glacier is very much emaciated, and the channels through which the water rushes are much wider. The current, while less violent, is still of such a character as to preclude any thought of navigation under any condition. Shooting the first and second rapids successfully the little party dropped down through the third and proceeded down the river past Childs Glacier. On the

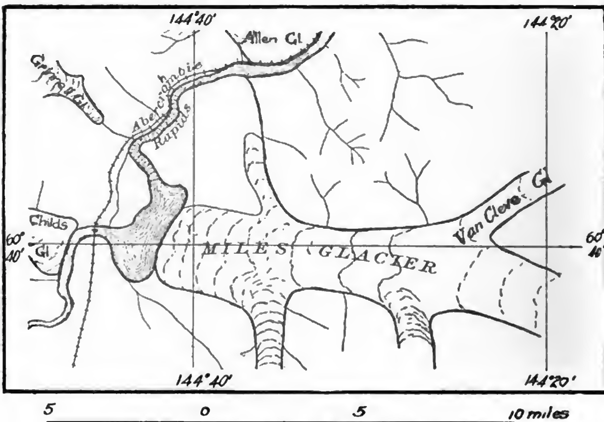


Fig. 61. Lower Portion of Miles Glacier in 1906.
After U. S. Geol. Survey.

faces of both Childs and Miles glaciers I noticed a very marked change. Miles Glacier had receded toward Mt. St. Elias some 5 or 6 miles. Where in 1884 the Copper washed the face of this glacier was now an immense lake.[1]

He further states that at the head of Abercrombie Rapids, Miles Glacier is "covered with an immense amount of morainic matter and a dense growth of alders, which effectively protect the glacier proper from inroads by the river." What has since been named Abercrombie Canyon "has a canyon wall on the right side only. The left is formed by the encroachment of Miles Glacier." The rapids are "known, respectively, as the first, second, third, and fourth rapids. When I first visited this canyon, in 1884, there was but one rapid, the first. On my return this season the emaciation of the

[1] Abercrombie, W. R., Reports of Explorations in the Territory of Alaska, 1898, War Dept., Adj.-Gen. Office, No. XXV, Washington, 1899, pp. 315, 320–322; Koehler, R. A., Same, p. 431; Rafferty, J. J., Same, pp. 449–450; also in Narratives of Explorations in Alaska, Washington, 1900.

glacier was very marked. To the left of the second rapid the river had cut out the bowlders and worn a huge hole in the glacier,[1] some 50 or 60 acres in area, which is the entering wedge for the formation of a new channel. The face of Miles Glacier below the fourth rapid, has receded some 5 or 6 miles to the east, or back from the river, and during the months of June, July, and August, the falling of bergs cut out by the comparatively warm river water from this glacier keep up a continual roar, as they fall off."

Corporal Koehler states that "during October tons and tons of ice broke off Miles Glacier which caused a noise like thunder and made the swells of the river almost reach the camp. In the morning the bay was found to be blocked by ice. Only to the extreme right was there an opening sufficient through which to navigate the boat." Guide Rafferty indicates that Miles Glacier was much the same in July as in October, 1898.

By October, 1900, the southern half of Miles Glacier seems to have retreated still farther, the ice cliff photographed by Schrader and Spencer and mapped by Witherspoon being farther back at least than in Hayes' sketch map of 1891, and the lake being larger.[2] This map does not represent as ice the stagnant northern portion of the glacier which causes Abercrombie Rapids, showing only a narrow distributary ice tongue north of the main glacier, doubtless where there was clear ice visible. Earlier observations prove that ice must have underlain this whole moraine-covered belt in 1900. A. C. Spencer has stated[3] that marked recession of Miles Glacier took place between 1899 and 1900.

In August, 1905, the Miles Glacier was photographed by Webster Brown, and in October, 1907, by F. H. Moffit, of the U. S. Geological Survey. The photographs by Moffit show that one lobe of the ice front terminating in the lake retreated approximately 1700 feet between 1900 and 1907, by comparison with the Schrader and Spencer photographs.

Observations by the Railway Engineers. The railway and bridge engineers of the Copper River and Northwestern Railway surveyed their right of way west of Miles Glacier and built their great steel bridge and the railway between 1906 and 1910, mapping the cliff of Miles Glacier, the lake and its outlet, and the Abercrombie Rapids. Their maps show marked enlargement of the lake between 1900 and 1908, with a slight recession on the south side of the glacier so that the middle projects in a pronounced point. A later map shows advance of the glacier and decrease in width of the lake.

The railway engineers made soundings in the lake, measured the velocity of the stream at the outlet of the lake, determined the rate of iceberg flow, bored through the glacial deposits near the bridge, kept a detailed meteorological record for three years, and made many general observations, from their camp between Childs and Miles Glaciers, where they were building the steel bridge across Copper River and operating a car ferry across the Miles Glacier lake in 1909 and 1910. Many excellent photographs of Miles Glacier were also taken by individual engineers, and by E. A. Hegg, of Cordova.

[1] The Bearhole.

[2] Geology and Mineral Resources of a Portion of the Copper River District, House Doc. 546, 56th Congress, 2nd Session, Washington, 1900, Pl. XIII B, and map, Pl. II; also reproduced as Pl. XII, in Bull. 284, U. S. Geol. Survey, 1906. A later edition of Witherspoon's map of Miles Glacier, with slight modifications, including the widening of the lake through the retreat of the ice front, has also been published by the U. S. Geol. Survey, Bull. 374, 1909, Pl. I.

[3] In H. F. Reid's Variations of Glaciers, Journ. Geol. Vol. IX, 1901, p. 253.

Observations by the National Geographic Society's Expeditions. Our own studies of Miles Glacier were made by both authors in August, 1909, and by the junior author in August and September, 1910, and in June, 1911. The first set of observations [1] was limited to parts of two days, while those in 1910 [2] occupied parts of two weeks, during which we made detailed studies and took many photographs, while Mr. Lewis made the topographic map reproduced as Map 9 (in pocket). The study of Miles Glacier in 1911 occupied only a few days, dealing chiefly with conditions of former glaciation rather than with Miles Glacier itself.

The Miles, a Bulb Glacier—General Conditions. Miles Glacier is an excellent example of a bulb glacier, a class intermediate between the valley glaciers and the piedmont glaciers. Miles Glacier is a bulb glacier which, on emerging from its mountain valley into the flatter grade of the broad Copper River valley, spreads out in a bulb or fan (Map 9). The valley glacier and the bulb will be discussed separately.

The Valley Glacier. The valley glacier (Fig. 61) appears to be perfectly normal. It has severely crevassed white ice with rather broad lateral moraines in which the ice is little crevassed. Its mountain walls (Fig. 62) rise to heights of 2300 to 4900 feet, with slopes at the rate of three to six thousand feet to the mile. The number of tributary

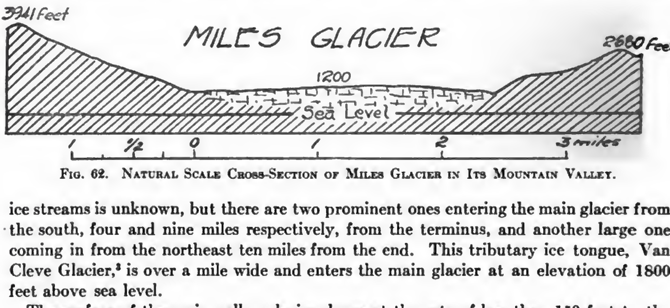


FIG. 62. NATURAL SCALE CROSS-SECTION OF MILES GLACIER IN ITS MOUNTAIN VALLEY.

ice streams is unknown, but there are two prominent ones entering the main glacier from the south, four and nine miles respectively, from the terminus, and another large one coming in from the northeast ten miles from the end. This tributary ice tongue, Van Cleve Glacier,[3] is over a mile wide and enters the main glacier at an elevation of 1800 feet above sea level.

The surface of the main valley glacier slopes at the rate of less than 150 feet to the mile, but three miles from the terminus, the slope steepens to 173 feet per mile, and just before reaching the ice precipice in the lake, the slope increases to 660 feet per mile, as if Miles Glacier valley had a hanging relationship to the main Copper River valley.

The Ice Cliff. The portion of the ice front which discharges bergs into Miles Lake is essentially the terminus of the valley glacier (Pl. CLIX). The southern edge is exactly at the mouth of the valley, for this part of the glacier does not spread out beyond

[1] Tarr, R. S. and Martin, Lawrence, Nat. Geog. Mag., Vol. XXI, 1910, pp. 13–14, 23, 37.

[2] Martin, Lawrence, Nat. Geog. Mag., Vol. XXII, 1911, pp. 542–543; in H. F. Reid's Variations of Glaciers for 1910, Journ. Geol., Vol. XIX, 1911, p. 458; Zeitschrift für Gletscherkunde, Band VI, 1911, p. 102; Mastering the Alaskan Glacier Barriers, Scientific American Supplement, Vol. LXXI, 1911, pp. 305–307; Petermanns Geog. Mitteilungen, Jahrgang 1912, Septemberheft, Tafel 9.

[3] Named in 1910, for J. R. Van Cleve, General Manager of the Copper River and Northwestern Railway.

the mountain valley; but the northern edge merges into the moraine-covered piedmont bulb. The ice cliff is irregular, with great capes and coves, but it is not as high as the cliff of Childs Glacier. The form of the ice cliff has varied in detail throughout the period of observations and there have been several changes of moderate amount.

The Bulb and the Marginal Channel. The second or bulb portion of Miles Glacier is represented by the stagnant northern part, and by the lake basin from which the bulb has melted away. The stagnant northern part of the bulb still extends clear across the valley, as it did at the time of the last great advance when Copper River was forced over into the marginal channel (Pl. CLVIII, A) which it now occupies. This northern portion, which rises to a height of 430 feet above the lake, or 500 to 555 feet above sea level, is all covered with ablation moraine, on which is a continuous alder growth, except where, as the ice melts, slumping has undermined the vegetation and let the moraine slide off the ice. Many small patches of this nature opposite Abercrombie Rapids were seen by us in 1909–11 and some also show in the woodcut from Abercrombie's 1884 photograph. The river has been held in the marginal channel through lack of retreat of this portion of the glacier, which is of course much retarded by the mantle of ablation moraine and vegetation. Many enormous glacial bowlders skirt the river bank next the glacier, evidently helping to keep the river in its channel. At one point a cove, known locally as "The Bearhole," was formed in the glacier by slumping between 1884 and 1898.

When the glacier melts completely out of Copper River valley the site of the rapids will be marked by an abandoned marginal channel, perhaps cut somewhat into the rock, perhaps with one side missing, where the ice-wall now rises. It is, however, possible that in the meantime the river may so entrench itself in a rock gorge at Abercrombie Rapids that it will be unable to regain the middle of the valley after the stagnant northern part of the glacier melts away. It is not known how rapidly the river is cutting, but it flows with great velocity (Pl. CLX), is heavily charged with sediment, and receives additional load from the rock and ice walls. At the foot of the rapids it has carried away all of the moraine left by the melting ice except a few of the largest bowlders.

The stagnant northern portion of the bulb has remained in about the same condition from some time before 1884 to the present time.

There are three quite different zones in the moraine-covered bulb which were studied in detail in 1910, and their relative areas represented on the map. The characteristics of these three parts are stated in the following paragraphs:

Zone of Thickest Ablation Moraine. The western or outermost part of the bulb (Thickest Ablation Moraine, Fig. 63) is deeply covered with ablation moraine, well seen in the portions facing Abercrombie Rapids (Pl. CLX). This moraine supports a dense growth of vegetation (Pl. CLXI) with trees estimated to be about 50 years old, and with only scattered small areas where the underlying ice is revealed by slumping; but ice was still present in 1910, close to the bank of the Copper River at Abercrombie Rapids (Pl. CLVII), and along the northern border of the bulb opposite Allen Glacier. Abercrombie's photograph, taken 26 years before our visit, shows as dense a growth of vegetation on this portion of the bulb as at present.

Zone of Thick Ablation Moraine. The second zone (Pl. CLXV) is also deeply buried by morainic débris underlain by ice. Here the ice is inferred to be less thickly covered

than in the western zone, because it is revealed in more abundant sections and is slumping more continually, as is shown by the general absence of vegetation. The line between these two zones is a fairly sharp one in some places, with an east-facing bluff between the forested western zone of dense vegetation and the relatively-barren zone east of it. There are a very few scattered shrubs, mostly alders, which appear to be not over a year or two old, for the largest were only a foot or two in height, and the stems but a quarter of an inch to an inch in diameter, but those that we cut down had from 14 to 22 annual rings. The stunted condition may be due either (a) to the glacier wind and the coldness of the soil, or (b), as we consider more likely, to the shifting of the soil with slumping, as the ice underneath melts, killing or retarding the growth of the moss, grass, and shrubs. This process was actively in progress during our visit in 1910.

In this zone, especially near the western edge, there were large depressions in the ice, with a depth of from 30 to 100 feet, often containing ponds. At their edges the moraine, which was 2 to 3 feet thick, was continually sliding down into the water, and the ice thus revealed (Pl. CLXII) was black with débris, clearly indicating the source of the material making up the ablation moraine. There were also a few long, narrow valleys without lakes, one of them 150 feet or more in depth, with steep walls of bare ice, suggesting the enlargement of a great crevasse. None of these valleys was continuous for a great distance. Farther east there were many débris cones, rising 15 to 25 feet above the general level, some of them made up of till with striated, subangular bowlders, and containing no ice. Many of these débris cones occur along lines, forming discontinuous ridges of distinctive rock, sometimes differing from the general rock material of the ablation moraine, sometimes composed of coarse stony material, rising slightly above the general level of an area of finer débris. These rows of débris cones are interpreted as the sites of former crevasses.

Three pronounced ridges close to the Copper River, near the southern border of the zone of thick ablation moraine (near Photo. Sta. M, Map 9), were from 10 to 15 feet high, about 100 feet apart, and made up of very coarse, angular rocks. There was slumping in one of these between 1900, when Schrader took a photograph, and our visit in 1910, showing that this area is still underlain by ice. Whether these three ridges are crevasse deposits, or recessional moraines is uncertain.

A narrow strip of moraine on the bank of Copper River and at the extreme southwestern point of the area of thickest ablation moraine, does not seem to be slumping and may contain no ice at present. There are large ponds in it and it apparently has not changed at all between 1884 and 1910.

The northern margin of the river and lake in the zone of thick ablation moraine is made up of (a) moderately steep cliffs of ice from which the débris is sliding, (b) partly of sand deposited by the Copper River and made into a broad sloping beach by the iceberg waves and (c), in the part bordering the channel by which the main river enters the lake, of very coarse stream-rounded bowlders.

Zone of Thin Ablation Moraine. In the third zone (Thin Ablation Moraine, Fig. 63), the ablation moraine forms an exceedingly attenuated veneer in most places, and bears no vegetation whatsoever (Pl. CLXIII). It is evidently an expanded extension of the northern lateral moraine of the valley glacier, and is separated from the barren zone of thick ablation moraine by a well-defined line, a low, west-facing cliff 5 or 10

feet high, with the zone of thin moraine standing distinctly higher than that of barren thick moraine. The eastern edge of this zone is at essentially the same level as the clean ice of the valley glacier.

Cause for Moraine Zones. The separation of the northern portion of Miles Glacier bulb into three distinct zones calls for explanation, for the condition is distinctly not that of gradation from zone to zone, but of sharp contrast (Pl. CLXIV). In the table below the significant facts are contrasted.

TABLE CONTRASTING CONDITIONS IN THE THREE ZONES OF ABLATION MORAINE ON MILES GLACIER

Zone	*Moraine Cover*	*Slumping*	*Vegetation*	*Age of Trees*
Thickest moraine	Several feet	Very little	Continuous	50 to 75 years
Thick moraine	A few feet	A great deal	Scattered	14 to 22 "
Thin moraine	A few inches	Rapid melting	None	....

Fig. 63 is a cross-section showing the three zones graphically. The east-facing bluff at b, and the west-facing ice cliff at c, as well as the marked difference in ages of trees, are thought to be especially significant in connection with the problem as to the difference of conditions in the forested zone (a–b, Pl. CLXIV), the relatively-barren zone (b–c), and the absolutely-barren zone (c–d).

At c, along the contact of the zones of thick and thin moraine, an advance was in progress or was just ceasing in 1910. At the edge of the lake it was clearly still in progress, for ice blocks were sliding down from an ice cliff which was not undercut by the

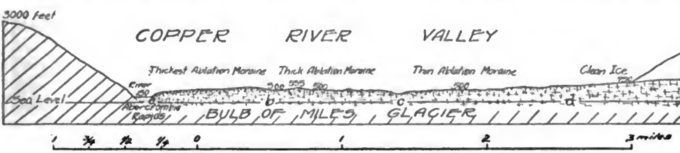


FIG. 63. NATURAL SCALE CROSS-SECTION OF COPPER RIVER VALLEY.

waves, and they lay in a heap on the beach. Back from the lake shore the snow of the winter of 1909–10 was folded up in front of the advancing ice edge. While we could not be certain that advance was still taking place all along the low cliff on the line c–c (Pl. CLXV), it was certainly still in progress for a quarter of a mile or so from the lake shore. The ice in the zone of thin moraine had the appearance of being thrust-faulted forward over the ice in the zone of thick moraine.

The explanation that naturally suggests itself is that an advance of Miles Glacier a few years before 1910 crevassed the ice in the zone of thin ablation so that the vegetation which had previously straggled across into this belt was destroyed, while the gradually-thinning morainic cover was partly swallowed up in the crevasses. This advance, of which there is independent photographic evidence, began after 1908 and had ap-

parently almost ceased in 1910 along the northern portion of line c and was just dying out near the lake shore. Melting had nearly healed the crevasses (Pl. CLXIII), and the removal of the upper layers of ice by ablation had left a thin layer of morainic débris upon the ice in the easternmost zone.

If this explanation is correct, it naturally suggests the possibility of a similar explanation for the difference of moraine and vegetation in the middle and western zones, whose sharp contrast on either side of the line b–b calls for a special explanation. There is independent evidence that Miles Glacier had an earlier and greater period of expansion. This evidence comes from the southern margin of Miles Glacier where there is moraine and outwash with two independent sets of vegetation 20 and 70 years old respectively. It is, therefore, thought that during an advance 22 years or more before 1910 there was expansion of the northern bulb of Miles Glacier out to b. This resulted in crevassing and destruction of the vegetation which had presumably grown in the zone of thick ablation moraine, similar to the breaking of the ice and destruction of the vegetation which in 1906 took place in a part of the piedmont bulb of Atrevida Glacier. Certain faint ridges in the zone of thickest moraine suggest the former presence of crescentic cracks, similar to those that developed in the outer part of the forested zone on Atrevida Glacier, outside the zone of complete destruction of the morainic mantle and its vegetation. The rapid slumping, the presence of many kettles with pools, the youthful vegetation, and the lack of gradation features at the border (b–b), all accord well with this explanation. The difference between the bordering lines at b and c seems to be merely one of time. The uppermost portion of the middle zone still rises higher than the western zone, but at the border the relatively cleaner ice of the area of thick ablation moraine has melted down below the level of the very dirty, protected ice of the zone of thickest ablation moraine. In similar manner the west-facing cliff at c should, in time, be converted to an east-facing cliff, because of difference in rate of melting in the areas of thin and of thick moraine, unless continued advance in the zone of thin moraine makes up for the losses by ablation.

Western Terminal Moraine. On the western bank of Copper River, between Grinnell Glacier and the mouth of Chinaman Slough, east of the railway bridge, there is a terminal moraine which seems to mark the westernmost extension of the Miles Glacier bulb.

Near Grinnell Glacier this terminal moraine forms a line of irregular hills extending nearly north and south between the railway and the northern part of Chinaman Slough (Pl. CLXV). Mature cottonwood trees grow on this moraine, which is less than an eighth of a mile wide and 30 to 50 feet high. Southward the moraine disappears beneath outwash gravels, reappears west of the railway and crosses the track north of Mile Post 50, and again disappears near the southern part of Chinaman Slough. But from here on to the site of the north ferry slip on the lake shore, its position may be traced practically all the way by the great numbers of large bowlders, some of which are from 10 to 12 feet in diameter. We believe that this belt of huge stones represents the residue of the terminal moraine after the river had washed away the clay, sand, and smaller bowlders. West of this moraine there are outwash gravels.

The Lake Basin. The site of the southern half of Miles Glacier bulb, east of the terminal moraine just described, is occupied by the lake basin, and the contiguous narrow zone of ground moraine from which the ice has completely disappeared. This

represents an area in which the ice has been retreating rapidly in the period since before 1888, presumably because the valley glacier is not moving forward strongly, is not protected at the terminus by morainic débris, and is undercut by the water. The site of the lake may possibly be classed as analogous to the interior flat areas observed in other bulb glaciers.

In 1910, the lake was from one to two miles wide, 3¾ miles long, and 150 to 300 feet deep (Pl. CLXV). Since the low-water surface is 116 feet above tidewater, the bottom, therefore, descends to 184 feet below sea level (Fig. 64). So large a lake in the course of a heavily-loaded river, and particularly of a glacial stream like Copper River, is an unusual feature, for such sediment-laden rivers rapidly fill and destroy lakes. The presence of the lake is of itself clear evidence that the glacier has recently receded from the site of the lake basin. The great depth of the lake is evidence along the same line, for at the mouth of Copper River the deposits have built up a large delta extending out into the Pacific Ocean, and it is necessary to go over 20 miles offshore before a point in the ocean is reached with a depth as great as that of Miles Lake.

Southern Terminal Moraine. South of the lake is a strip of valley bottom a quarter to a half mile wide which also represents parts of the former bulb of Miles Glacier.

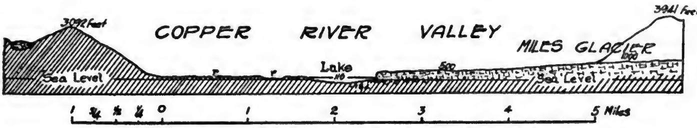


FIG. 64. NATURAL SCALE LONGITUDINAL SECTION OF MILES GLACIER IN 1910.

This is proved by the crescentic terminal moraine which swings across the Copper River valley from the eastern valley wall a half mile south of Miles Glacier to the railway bridge at Miles Glacier station (Pl. CLXV). This moraine is 70 to 90 feet higher than the outwash plain on the south and 20 to 30 feet higher than the area of ground moraine which separates it from the lake. One kame hill in this moraine rises to a height of 186 feet above the low-water surface of the lake.

Most of the moraine is a series of discontinuous ridges forming a belt varying in width from 100 feet to a quarter mile, and with a general trend parallel to the former ice front. In the kettles between these ridges are a number of ponds.

Ground Moraine and Detached Ice Mass. Between the terminal moraine and the lake is a parallel strip of ground moraine and one large, detached ice mass. The ground moraine is very irregular and has many deep kettles containing pools, and a rolling surface of low ridges, standing 60 to 100 feet higher than the outwash plain south of the moraine. It is made up of till and gravel, and, in large areas, of angular rocks of considerable size (Pl. CLXVI), sometimes arranged in long lines of crevasse deposits.

The edge of the detached ice mass south of Miles Glacier is buried beneath the outwash gravels which mantle part of the ground moraine, and these gravels were slumping in 1910. The detached ice mass is riven by great crevasses, and one of these, 50 to 90 feet deep, had morainic débris sliding down into it (Pl. CLXVII, A) from the thick covering of ablation moraine resting on the ice. When the ice has completely

melted away, there will be a conspicuous ridge of rocky material of the sort interpreted elsewhere as crevasse deposits.

Where the ground moraine forms the southern shore of the lake there are several capes and bays, and large valleys and kettles, apparently corresponding to former lobes in the ice edge. Here cliffs rise 40 or 50 feet above lake level, and the slope continues steeply below water, as is shown by the fact that in one place 375 feet from the shore the lake is 150 feet deep, while in another place it is 45 feet deep 50 feet from shore, indicating that these are ice-contact slopes rather than wave-cut cliffs, though above lake level they have been somewhat modified by wave work.

Outwash Plain. The southern edge of the terminal moraine is fringed by a series of small alluvial fans of coarse gravel leading out of little ravines in the moraine. A few hundred feet from the moraine they merge into a smooth, gently-sloping outwash plain which extends southward and westward. This plain is trenched by scores of sub-parallel, southwest-trending, abandoned stream courses which form the only irregularities in the outwash plain, with the exception of two groups of hills of older moraine.

The border between the moraine and the outwash plain is generally sharp, but in places the moraine is so low that only the nature of the soil indicates the boundary. This is true for a short distance southeastward from the railway bridge where the till and large sub-angular bowlders of the moraine, in contrast with the assorted rounded gravel and smaller stones of the outwash plain, are more useful than the topography in mapping the moraine.

Vegetation on Moraine and Outwash. The vegetation in this southern area formerly occupied by the Miles Glacier bulb, covers the outwash plain, the terminal moraine, and part of the ground moraine. On the outwash gravel plain there is a thick growth of cottonwood and dense underbrush, with trees a foot or two in diameter, the oldest one we found, a cottonwood, having 70 annual rings. In the abandoned stream channels which extend across the outwash plain, there are no trees as old as those between the stream courses. The alluvial fans at the base of the terminal moraine are covered with moss and grass but have no mature forest like that on the outwash plain.

The terminal moraine is about as densely covered with vegetation as the outwash plain, but there are no cottonwoods. The woody plants are all alders and willows and they form such a complex network of close-set, inclined, interlacing trunks that travel through the thicket is very difficult. The plants are younger than those on the outwash plain, the oldest, whose age we determined, having 20 annual rings.

In the belt of ground moraine the vegetation is progressively younger along a line from the terminal moraine along the lake shore, toward the ice edge the ages of the bushes decreasing from 19 to 5 years. Acres of moraine, too rocky for the rapid growth of bushes, were still absolutely barren in 1910, but wooded slopes extend out to the lake shore, and an alder with 19 annual rings was found within a quarter mile of the glacier. There were no shrubs near the retreating edge of the detached ice block, but some small shrubs were growing in the ablation moraine upon its surface.

Evidence of Two Stages. The difference in ages of trees on the moraine and on the outwash plain calls for a special explanation. It is impossible that the ice stood at the terminal moraine up to something over 20 years ago, while the vegetation upon the outwash plain was growing freely, for in that case the trees upon the outwash plain should have been buried in gravels, uprooted, or killed by the water.

It accords well with the known history of the northern portion of Miles Glacier bulb to infer that when the glacier advanced across the Copper River valley to the site of Abercrombie Rapids, it also expanded southward to or near the site of the present terminal moraine, but probably not beyond it. This was at least 70 years ago (that is, before 1840), and it may well have been still earlier. In the southern section the ice has retreated from this stand, though it is still maintained in the zone of thickest ablation moraine in the northern portion of Miles Glacier at Abercrombie Rapids. The 70 year old trees on the outwash gravel plain indicate a similar expansion at the same period in the southern area.

A more recent advance, to be correlated with the one which extended across the zone of thick ablation moraine north of the lake, extended out to the present terminal moraine south of the lake, destroying any older moraines and vegetation between the lake and the moraine. After the terminal moraine was built the glacier again receded and vegetation sprang up on the surface. This was over 20 years ago, that is before 1890. The 22 year old shrubs north of the lake show that it was before 1888, but whether it was before the visits of Abercrombie and Allen in 1884–85, is not certain. Hayes' map made in 1891 and the 19 year old shrubs within a quarter mile of Miles Glacier, indicate that the retreat from this maximum was rapid, though it is to be noted that the single 19 year old bush may have had its start on the ice and have maintained growth during the melting of the ice beneath the moraine in which it grew.

The lack of destruction of all the older vegetation on the outwash plain suggests that the ice stood at the terminal moraine a very short time. The streams from the ice did destroy the vegetation in the border of the outwash plain, where the small alluvial fans form a rather barren fringe in front of the terminal moraine. They also cut swaths through the forest, now represented by the abandoned southwest-trending stream courses, in the bottoms of which we saw no 70 year old trees. These channels are cut so deeply below the general level of the plain that they suggest streams less heavily laden than those which deposited the outwash gravels. The channels are now overgrown with dense alder thickets, but we did not determine the ages of any of the shrubs in them. The absence of great numbers of dead cottonwoods standing in the stream courses seems to be due to the depth to which the channels were cut and the consequent removal of the tree trunks and roots; but some wood lies partly buried in these channels.

Scattered through the undergrowth, in the interstream areas, where the 70 year old cottonwoods grow, are a very few dead trees, standing upright in place, and a great many recumbent logs, evidently left by floods at the time of the cutting of the channels. That all the trees were not killed is an evidence of the shortness of the period of stream invasion of this forest. The short distance which the ice front had tore treat before the water from its melting would be diverted through the deep lake basin and find an outlet into the present channel of Copper River, at or near the railway bridge, doubtless helps to explain the lack of further destruction of the vegetation on the outwash plain during this episode.

An exceedingly detailed map of the portion of the moraine and outwash plain near Miles Glacier Station was made by the railway engineers. It has a contour interval of 2 feet and shows numerous shallow kettles due to slumping of ice formerly buried beneath the outwash gravels at the edge of the moraine; and it also shows some of the abandoned stream channels leading southwestward toward the Copper River. A conspicuous

abandoned channel of this series heads in the air just west of Miles Glacier station, proving that the Copper River has undercut its south bank since this small stream channel was abandoned.

History of Oscillations of Miles Glacier—Advance Before 1840. Passing over the early stage when Miles Glacier was a tributary of a great Copper River Glacier, and the legendary stage when Miles and Childs Glaciers are supposed to have coalesced, with Copper River flowing underneath, the first stage of which we have knowledge is the expansion to the bulb glacier condition some time before 1840. It is quite possible that this expansion was preceded by one of shrinkage of Miles Glacier back within its mountain valley, and that it expanded into a trunk valley which was free from ice. The asymmetrical shape of the bulb suggests that Childs Glacier also had an expanded bulb, for the Miles bulb spread 2½ miles up the main valley and, so far as we know, only a fifth as far down Copper River valley, forming a striking contrast with the far more symmetrical bulb of Allen Glacier (Map 9).

We know nothing specific as to the date of this expansion to the present bulb form excepting that (a) Abercrombie's photographs prove that it took place before 1884, (b) Grewingk's description (p. 416) suggests that it took place long enough before 1850 for vegetation to grow on the ice, and (c) the age of trees growing on the outwash plain pushes the date back to some time previous to 1840. It may have been much earlier.

Slight Recession During the Eighties. Maintenance of an advanced position rather than pronounced recession is the impressive feature of the history of Miles Glacier during the time of the visits by Abercrombie and Allen. The northern portion of Miles Glacier bulb has retreated practically none in over 70 years; and the southern portion also maintained its advanced position for a time, the site of the lake basin being largely filled with glacier ice until very recently.

Abercrombie's mention of a lake basin shows that there had been some recession of the southern part of the glacier by 1884, and one of his photographs,[1] taken from a point on the western terminal moraine of Miles Glacier, south of Grinnell Glacier, proves that the ice cliff had retreated to the point where the largest channel of Copper River now enters the lake. This position of the ice front represents over a mile of eastward recession from the expanded edge of the bulb at the southern end of Abercrombie Rapids. The great waves in the stream in the foreground of the same photograph show that a main channel of Copper River flowed down Chinaman Slough as recently as 1884. That the southern margin of Miles Glacier was not severely crevassed is indicated by the fact that Abercrombie traveled 8 miles upon its surface in August, 1884, being turned back by the crevasses in the middle of the glacier after traveling up its southern border.

In 1885 the front of Miles Glacier still filled the greater part of the present lake basin, as is indicated by Allen's descriptions, one of which includes the specific statement [2] that "the most southerly point of the . . . (Miles Glacier) is 1 mile or less from the most northerly point of the . . . (Childs Glacier)."

It is clear, therefore, that while there was slight recession of the southern portion of the Miles Glacier bulb during the eighties, it did not uncover a significant part of the present lake basin. A comparison of Abercrombie's 1884 photograph of the northern portion of the bulb with our photograph from the same site shows practically no change in the 26

[1] Reproduced by H. T. Allen in Narratives of Explorations in Alaska, Washington, 1900, picture facing p. 426.

[2] Allen, H. T., in Narratives of Explorations in Alaska, Washington, 1900, p. 486.

years up to 1910, a remarkable maintenance of a stagnant ice mass due to the protective influence of ablation moraine and vegetation.

Advance Before 1888. The ages of trees on the southern terminal moraine and in the zone of thick ablation moraine on the northern lobe make it clear that at some time over 22 years before 1910, there was a period of advance, breaking up the ice on the northern bulb to within three quarters of a mile of Abercrombie Rapids and building the southern terminal moraine. That this took place after 1884 and 1885 is suggested by Abercrombie's photograph and Allen's map which, though generalized, shows no southward extension of the bulb. This expansion of the southern margin to within 400 feet of the site of the railway bridge clearly came after 1885 and before 1888, and may have begun in the former year, for Allen's map shows the glacier nearly out to the bridge site. If this advance occurred in 1885 or 1886, it was exceptionally rapid, since in 1884, Abercrombie was able to travel up the southern margin of the glacier, and in April, 1885, Allen noted that the drift graded indistinguishably into the ice of the southern edge of the glacier.

Recession Up to 1908. A general retreat began by at least 1888 and had progressed so far that the lake was about a mile wide at the time of Hayes' visit in 1891.

The 5 or 6 mile retreat mentioned by Abercrombie on his second visit to Miles Glacier in 1898 is probably an exaggeration, for his map does not indicate that the lake was over a mile wide. This map does not show the details as well as Hayes' map and may represent the condition in 1884 rather than that in 1898. There was doubtless marked retreat, however, by 1898.

By 1900, when Schrader and Spencer each photographed the ice front on different dates in October, and Witherspoon made his map, the retreat of the ice cliff had been considerable. A careful comparison of the 1900 photographs with the conditions in 1910 indicated that the ice front had then receded so far that the lake had assumed nearly its present width, that is, different parts of the glacier front had retreated from 1½ to 2 miles from the western terminal moraine of the southern part of the bulb.

From 1900 to 1907, when Moffit visited Miles Glacier, retreat went on rapidly, different portions of the ice cliff melting back 1700 to 3900 feet during the 7 years interval. This recession was carefully determined in 1910 in two ways. First, we occupied the photographic stations of Schrader and of Moffit with their photographs in hand and made instrumental observations of the amount of recession, by projecting the glacier termini against the mountain wall behind. This retreat is shown graphically on the photograph (Pl. CLXVIII), being marked on the lobe which retreated 1700 feet. The second method of determination was by comparing maps made in 1900 by the Geological Survey and in 1906 by the railway engineers, based on surveys in 1905 and 1906. These maps show that while the lobe mentioned above was retreating 1700 feet during the 6 years from 1900 to 1906, another lobe, farther south, receded nearly 3900 feet. The larger part of the retreat was between 1905 and 1906.

That there was additional recession of 1620 feet between 1906 and 1908 is shown by another map of the railway engineers. The position of the ice cliff at this stage is shown in a photograph by E. A. Hegg, from the western terminal moraine.

Advance in 1909 and 1910. The change from retreat to advance came in 1909 and 1910 when there was a forward movement of over 4000 feet along one of the lobes. This was determined with some precision, for the railway engineers mapped the whole lake front

of the glacier in 1908 and again in June, 1910. On one photograph of the 1910 ice cliff from Schrader's site, we have indicated by an ink line the amount of advance between 1907 and 1910. This, however, shows only the lobe which retreated 1700 feet between 1900 and 1907, for the lobe which advanced 4000 feet is not seen in the picture because it is hidden behind the nearer lobe, having retreated farther before the recent advance.

There was slight additional advance of the ice cliff in the lake during the time of our visit in the summer of 1910, when between June and August the advance amounted to 100 feet. The southern glacier margin was crevassed and advancing in August, 1910, and the northern margin, which was also advancing in August, 1910, showed an appreciable forward movement between the times we photographed it in 1909 and 1910.

The retreat from 1888 to 1908 and the subsequent advance are summarized in the following table. The possible error in any given map would modify these amounts only slightly.

TABLE SHOWING AMOUNTS OF THE MOST RECENT RETREAT AND ADVANCE OF MILES GLACIER

Date		*Distance from Site of Railway Bridge to Ice Cliff*	*Change in Glacier*	*Amount*	*Based upon Observations by*
After 1885 and before 1888		400 feet	Advance	——	Henry T. Allen[1]
October,	1900	7,900 feet	Retreat	7,500 feet	D. C. Witherspoon
Fall,	1905	8,250 feet	Retreat	350 feet	Railway engineers
Summer,	1906	11,880 feet	Retreat	3,630 feet	" "
Summer,	1908	13,500 feet	Retreat	1,620 feet	" "
June,	1910	9,500 feet	Advance	4,000 feet	" "
August,	1910	9,400 feet	Advance	100 feet	Martin

During the advance in 1909 and 1910 the form of the ice cliff changed in detail (Fig. 65), though it is reported that twice as many icebergs were discharged in 1908, when the glacier was retreating, as in 1909, when it advanced.[2] One remarkable feature was the shifting of position of the salient ice point. There were also pronounced changes in the capes and coves of the ice cliff between June and August, 1910.

The most striking change was the detaching of huge masses from the glacier, one near the northern, the other near the southern margin of the ice cliff. These were not icebergs but large areas of the glacier itself, and their separation from a glacier which was advancing rather than retreating was an unusual circumstance.

About the middle of July, 1909, a section of Miles Glacier at least half a mile square, became separated and is said to have floated out into the lake, where the mass collapsed and broke up into icebergs which completely filled the lake. This was near the northern end of the ice cliff. It took place before our visit in 1909 and we have this information

[1] Date 1885 from Allen. Date 1888 from ages of trees as determined in 1910.

[2] Based on a count by A. O. Johnson at the railway bridge.

from Mr. A. O. Johnson.[1] In 1910 the gap made by this ice mass was still visible and near it there was a broad valley upon the glacier surface and this part of the ice cliff sloped

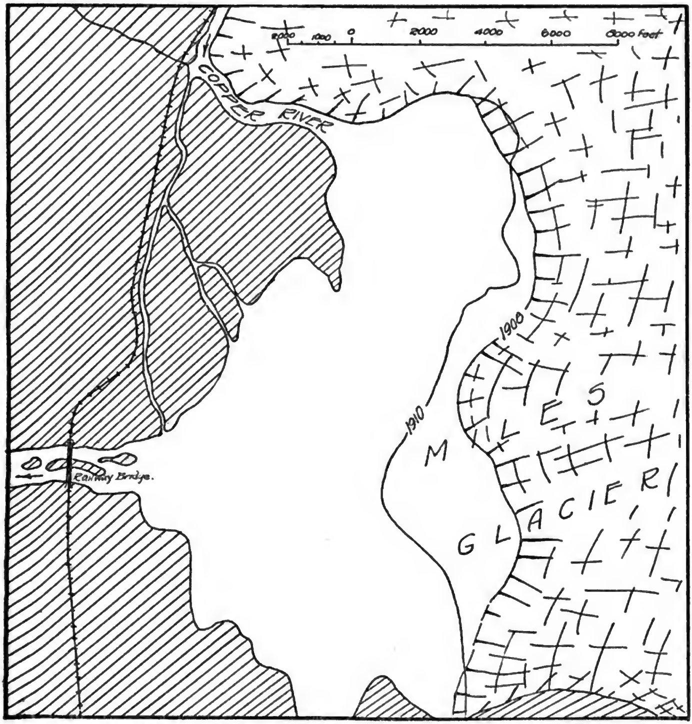


Fig. 65. Map Showing Advance of Ice Cliff of Miles Glacier in the Lake Between 1908 and 1910. (Compiled from maps by railway engineers.)

gradually to a 10 to 15 foot cliff at the water's edge instead of forming a precipitous terminus 200 feet high.

The cause of this change seems to go back to the winter of 1909, when there was a flood (February 9 to 18), during which Copper River rose 7 feet higher than any previous

[1] Personal communication, February 12, 1910.

record, the level going up over 21 feet above normal winter level, and falling rapidly again. Johnson states that the water came from beneath Miles Glacier at the point where the ice mass floated out 5 months later, and that "from the time of the break the river continued to rise for four days, was stationary one day, and then receded at about the same rapidity as the rise. The river at the bridge crossing . . . (before this flood) had a cross-section area of about ten thousand square feet and an average velocity of two miles per hour, but during the highest stage of the river the cross-section area was about forty thousand square feet, and average velocity ten miles per hour. . . . Considerable ice fell from the face of the glacier at the point of the break during the flow, but not until the middle of July did any great amount break away—at which time a section at least one half mile square, caved, floated out and completely filled the lake, which covers approximately five square miles."

The climatic record kept by the railway engineers at Miles Glacier bridge, shows no rainfall at or before the time of the flood and only the normal increase of snowfall for that season of the year. There were fluctuations in temperature before the flood began on February 9, but from the middle of December until two weeks after the flood the maximum temperature was not above the freezing point. The wind velocity did not increase abnormally immediately before the flood, though on the day when the water was highest the wind blew 48 to 84 miles an hour from the northeast. A high northeast wind in summer would drive the water from the place where the break occurred to the outlet of the lake but it could not have caused the initiation of the flood in February, for the lake was then frozen over, and the flood could not have kept up unless water was supplied from outside. The meteorological conditions, therefore, seem to have had little, if anything, to do with the flood.

Our own view of the detaching of the ice mass in 1909 is that it was not necessarily related to the advance or retreat of the glacier at all, but to undermining during the February flood which may have been caused by the draining of a marginal lake higher up Miles Glacier. A similar flood on August 16, 1912, perhaps from the draining of a marginal lake, swept down the Copper River from Miles Glacier. It raised the water level 12 feet at the railway bridge east of Childs Glacier and, 20 miles farther south, swept away 1600 feet of railway trestle east of Flag Point, drowning a repair crew foreman.

A second event of similar character occurred during the summer of 1910. At the time of our visit in 1909 and up to June, 1910, there was a stagnant, projecting edge of Miles Glacier which clung to the mountain side at the southern end of the ice cliff and merged into the glacial deposits there. Sometime between June 10 and August 15, 1910, a section of the glacier, measuring over an eighth of a mile on each side, floated away, leaving the outer stagnant ice as a detached block, 800 by 1200 feet, and extending up the mountain slope to a height of 175 to 200 feet above the lake. The whole of this stagnant block was thickly covered with ablation moraine, and it was riven by great cracks. To the southward it merged into slumping gravel deposits. Since the southern ice edge was still advancing after this separation took place, one possible method by which the separation may have been effected is the shove of the advancing glacier against the stagnant border which was clinging to the mountain side, thus placing a portion of it in such unstable equilibrium that it broke up.

In June, 1911, Miles Glacier showed no appreciable changes since the previous year.

In 1912 Corser [1] thought that it had advanced very slightly. We infer from this that the period of activity and advance which began in the fall of 1908 or the spring of 1909 was nearly or quite over in 1911.

Contrast with Advance of Childs Glacier. The advance of Miles Glacier in 1909 and 1910 was without spectacular features. There is little current in the portion of the lake near the ice cliff. The iceberg waves attack the ice cliff little, and melting by the lake waters, though undercutting the ice cliff as salt water does, has not undermined the cliff of Miles Glacier as rapidly as the Copper River mechanically undercuts the Childs Glacier. The infrequency of iceberg falls, and the great distance for the sound to carry, cause persons who traverse the Copper River railway to pay little attention to Miles Glacier, especially as the waves from it nearly disappear before they strike the shore near the railway bridge. This is why the pronounced advance of Miles Glacier in 1909 and 1910 attracted little attention on the part of the thousands of men who were along the line of the railway and the scores who were living near the railway bridge in these years.

The rates of movement during the advance of Miles Glacier were not measured. If it took the whole two years (from 1908 to 1910) for the ice front to move forward the

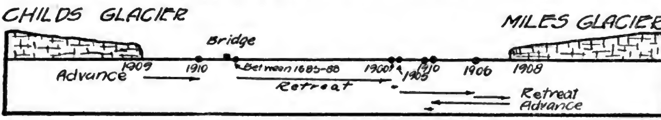


Fig. 66. Diagrammatic Sketch Showing Relationships of Miles and Childs Glaciers to the Railway Bridge.

The black dots in the horizontal line show positions of the ice cliffs in the years indicated, the advance and retreat between dates being indicated by the arrows below.

measured 4000 feet, the rate of motion was less than 6 feet a day, plus, of course, what was lost through iceberg discharge. From June to August, 1910, it was less than 2 feet a day. In any event it is clear that the recent advance of Miles Glacier was of a mild sort, compared with the rapid and far more spasmodic advance of Childs Glacier in 1910.

We have represented by a diagram (Fig. 66) the relative distances from Miles and Childs Glaciers to the railway bridge at several times in their recent history, selecting the bridge as the point of measurement, because it furnishes a convenient common point. It is clear that if the advance of the great Miles Glacier in 1909 and 1910 had been of the same order of magnitude as that of the smaller Childs Glacier in the latter year, the Miles ice front would easily have gone out to or beyond the railway. That such an event is possible is evident from the advance between 1885 and 1888, which took the ice cliff to within 400 feet of the site of the present bridge.

That this structure could not be placed better in its relation to both Miles and Childs Glaciers, is clearly indicated by the advances of 1888 and 1910. The facts suggest also that a possible great future advance of Miles Glacier would probably be short-lived. If the ice expanded again to the southern terminal moraine, the glacial streams might tear out or bury the railway grade for long stretches south of the bridge, but the deep lake basin and the present channel should soon dispose of most of the water, as they did at the last expansion. Besides possible destruction of the bridge during an advance,

[1] Corser, Caleb, Personal communication, September 30, 1912.

there is also a possibility of a southward shifting of the river bed, necessitating the building of another steel span.

From the point of view of tourists, who will undoubtedly make increasing use of the Copper River railway in the future, the location is ideal, for with no attendant danger or hardship, they can be carried by train almost up to the fronts of two very grand and active glaciers, both clearly visible from the train windows.

Grinnell Glacier

General Description. This ice tongue is on the western side of Copper River about half way between Childs and Allen Glaciers, terminating opposite the edge of Miles Glacier at the southern end of Abercrombie Rapids. It rises in unexplored snowfields at an elevation of at least 5000 or 6000 feet (Pl. CLXIX) and flows southeastward, with a known length of something more than 4 miles. It is over three quarters of a mile wide and has a moderately steep slope. The surface of the glacier is all white, except for a broad lateral moraine on the southern border, a narrow medial moraine near the northern side, and an extensive area of outwash gravel and ablation moraine covering the terminus (Pl. CLXV).

The real terminus of Grinnell Glacier is in the Copper River (Map 9), at the lower end of Abercrombie Rapids, just south of Camp 52 on the Copper River and Northwestern Railway. The railway runs over the end of the glacier for nearly a quarter of a mile, but the terminal hill is so thickly covered with ablation moraine, on which dense thickets of alder and cottonwood have grown, that this portion of the ice tongue does not look like a glacier.

The southern portion of the glacier, terminating less than a quarter mile west of the railway, is also heavily covered with ablation moraine and forest, which grades northwestward to a barren region of thinner, ablation moraine and to clear ice without morainic covering.

The northern terminus of the clear ice, which seems to be the glacier terminus, but is not, ends a little over half a mile northwest of Camp 52, on the railway. From it a stream flows through the interior flat and through a narrow gorge between the high outer portion of the glacier and the mountain side, to Copper River. Another stream, emerging from the southern part of the glacier near a conspicuous, bare rock slope, from which the ice is completely melted, flows southeastward over a mile to the Copper River.

Previous Observations. This glacier was mapped by Hayes in 1891[1] and Witherspoon in 1900,[2] when it was also photographed by Spencer from the Miles Glacier portage. It was photographed by Moffit in 1907 from near the same site as that occupied by Spencer, and was shown upon a later edition (Fig. 68) of Witherspoon's map.[3] Witherspoon's original map (Fig. 58) shows the glacier more accurately than the later edition in which the forest-covered terminus of the glacier is omitted. Hayes' map is more like the later edition of Witherspoon's but it has the name *moraine* printed between the glacier

[1] Unpublished map.

[2] Pl. II, facing p. 28, in Schrader and Spencer's Geology and Mineral Resources of a Portion of the Copper River Region, Alaska, House Doc. 546, 56th Congress, 2nd Session, Washington, 1901.

[3] Pl. I in Bull. 374, U. S. Geol. Survey, 1909.

terminus and the river, though without suggestion whether this moraine was underlain by ice or not. Hayes does not mention this glacier in the text.

Both Spencer's and Moffit's photographs show clearly that the stagnant outer portion had almost exactly the same relationships in 1900 and 1907 that they had in 1910. In 1906 and ensuing years incidental observations of Grinnell Glacier were made by the railway engineers, who named this ice tongue. In 1909 two important photographs of Grinnell Glacier were taken by E. A. Hegg of Cordova.

In 1909 we saw Grinnell Glacier in passing, but gave no time to its study. In August, 1910, and in June, 1911, four days were devoted by the junior author to its examination.

Cascading Ice Tongue. The snowfields of Grinnell Glacier are not visible from its lower portion, but the abrupt topography suggests that they occupy a series of deep glacial cirques. There are at least three important tributaries. The clear ice terminates in two minor lobes. Most of the clear ice portion of the glacier that we have seen, photographed, or mapped, is cascading down a moderate slope and is severely crevassed. The highest point on the glacier whose elevation we determined is 2000 feet above sea level or 1500 feet above the terminus of the clear ice, the slope here being at the rate of nearly 2000 feet to the mile. Higher portions of the ice tongue near the snowfield have even steeper slopes, but in an intermediate portion above the 2000 foot contour the grade of the glacier flattens somewhat. This cascading terminus is interrupted as the result of a discordance of over 2000 feet between the grades of Grinnell and Copper River valleys. The glacier still occupies its hanging valley and expands slightly in the moraine-covered terminus at its base. The thinness of the ice on the hanging valley lip is shown by the bare rock slope between the northern and southern lobes.

The mountain valley occupied by the glacier is exceptionally steep-sided, and also those of two detached tributary glaciers on the northern wall. These hanging glaciers now terminate 2300 and 3200 feet above sea level and their slopes are over 2800 feet to the mile. Just east of these hanging glaciers the northern valley wall slopes at the rate of over 6300 feet to the mile but the southern wall is not so steep.

Stagnant Southern Lobe. The southern lobe of the clear ice grades eastward through a zone of barren, thin, ablation moraine to a zone of thick ablation moraine, which mantles the ice so completely that it supports vegetation which entirely covers the surface. The end of the lobe does not look at all like a glacier and would not be known to be underlain by ice were it not for occasional slumping of the surface in the forested zone, and the exposed northwest-facing slope of black ice, studded with bowlders, in the zone of barren moraine.

Forested Hill at Terminus. Grinnell Glacier terminates in a forested hill (Fig. 67), which we believe to be wholly composed of ice. This belief is based upon the fact that here and there in this outer portion are areas of slumping where trees are being overturned. In several places, a whole steep slope has lost all its vegetation by the slumping attendant upon melting of the ice beneath. The cliff facing the edge of the interior flat at a number of points shows dark ice, filled with bowlders. At one locality near the northern edge the thickness of the ice revealed is 30 to 35 feet, above which are about 2 or 3 feet of morainic material in which shrubs are growing (Pl. CLXVII, B). Throughout the terminal hill of moraine-covered ice there are a great many areas of recent slumping and of emergence of water. The shrubs growing on the high outer

portion of the glacier cover the whole surface thickly, except in areas of slumping. They include alders at least 18 years of age.

The terminal hill rises from the water's edge in Copper River in a steep slope, west of which there is a long gradual ascent to an elevation of about 400 feet and a steep descent over a westward-facing cliff to the interior flat.

Much of the morainic material covering the ice of the terminal hill is angular and the bowlders are exceptionally large. They include many rock masses 12 to 20 feet on a side. Some of these huge rock masses in the railway cut on the eastern edge of the glacier are so large that they lead to the impression that they are rock ledges in place. This is not the case, however, as was ascertained by careful study in 1910. None of the other glaciers of the lower Copper River are encumbered with such large débris, except the western terminus of Miles Glacier at Abercrombie Rapids. This may possibly be explained by the exceedingly steep walls of the valley of Grinnell Glacier,

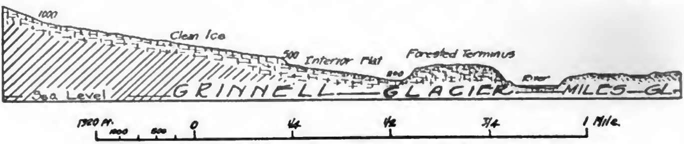


Fig. 67. Natural Scale Longitudinal Section of Grinnell Glacier in 1910.

upon which avalanches should be prevalent. As such coarse material is not found in the present lateral moraines of the cascading portion of Grinnell Glacier, however, it may be that the coarse angular débris was avalanched down upon the Grinnell Glacier during some great earthquake in the past and carried forward to the terminal bulb in the subsequent advance of the glacier.

The Interior Flat. Between the terminal hill and the cascading portion of Grinnell Glacier is an interior flat (Pl. CLXIX) between a quarter and a half mile square. Its surface is covered with ablation moraine and outwash gravel, overlying the ice, which is nowhere exposed. As Fig. 67 shows, the expanded part of Grinnell Glacier may possibly constitute a detached ice block, although it seems more likely that the ice is continuous beneath the ablation moraine and outwash, for areas of slumping are found here and there between the terminal hill and the clean ice of the cascading portion of the glacier.

The interior flat consists of three different portions. In the middle is a broad, low ridge of irregular ablation moraine. On each side of this are alluvial fans of outwash gravels.

The medial ridge rises 50 to 100 feet above the flanking alluvial fans. Its surface is made up of angular ablation moraine; the bowlders being decidedly smaller than in the terminal hill. The water which oozes out here and there shows that ice underlies the moraine at no great depth. Compared with the terminal hill this portion of the interior flat is comparatively barren, but small scattered shrubs are growing upon different portions of the surface. None of them are as old as those on the terminal hill.

North of this medial ridge is a detached ice block, which lies upon the northern valley wall. It is separated from the medial ridge by outwash gravels. Its surface is completely mantled by ablation moraine upon which a few small shrubs are growing.

The remainder of the interior flat is mantled with outwash gravels supplied by three main glacial streams, one north and one south of the medial ridge and one between which flows into a subglacial channel beneath the ice of the terminal hill. These gravel surfaces are almost entirely barren. Practically all of the northern area of outwash is within the interior flat, although the stream cuts through the terminal hill in a narrow, steep-sided gorge which ends at Camp 52 on the railway. The southern and larger area of outwash gravels also extends outside the interior flat and is continuous with a large alluvial fan southeast of the glacier. This stream emerges from the interior flat in a constriction between the terminal hill and the stagnant southern lobe of the glacier. These outwash gravel fans slope at the rate of 400 to 900 feet to the mile.

Recent History of Grinnell Glacier. The first stage in the recent history of Grinnell Glacier of which record is preserved is a time of sufficient expansion so that lateral moraine terraces were built on the valley sides several hundred feet above the present glacier. With this augmented thickness the glacier would naturally have extended slightly farther into the Copper River valley than at present. Whether it then was a tributary of the trunk glacier of the Copper River valley or merely touched the distal portion of the more powerful Miles Glacier cannot be determined. The relationships of the stagnant moraine-covered termini of Grinnell and Miles Glaciers (Pl. CLXV), make it clear that at some stage in the last period of great expansion these ice tongues, whose termini are still within 500 feet of each other, must have coalesced, the Copper River flowing beneath the ice or else maintaining a channel between the two ice tongues. This latter would be possible if the advances of Grinnell and Miles Glaciers were not synchronous, but the density and age of the vegetation on the two termini suggest that they were.

From 1891 to 1907 the Grinnell Glacier is not known to have changed appreciably. The photographs by Spencer and Moffit show no marked differences from the glacier of 1910. In 1900 there was a forested terminal hill and an interior flat. The age of trees on the terminal hill suggests that the last great expansion may have come shortly before 1892 but the observations of Hayes in 1891 and of Abercrombie and Allen in 1884 and 1885 furnish no details as to the previous conditions.

Mr. E. C. Hawkins informs us that in the autumn of 1907 there were great slides of morainic material from the terminal hill of Grinnell Glacier near Camp 52 and that a new stream appeared at that time.

A slight advance has been in progress from some time before 1907. This affects only the terminus of the cascading portion of the glacier. A comparison of the photograph by Spencer in 1900 with one by Moffit in 1907 indicates that a small area of the slope of bare rock in the middle of the cascading terminus was covered by the ice between 1900 and 1907. This advance may have continued ever since, or it may have ceased and a new period of advance started, for there was distinct minor advance from 1909 to 1911. Comparison of a 1909 photograph by E. A. Hegg, with the glacier terminus of 1910 showed marked advance. This continued at least up to June, 1911. It seemed to be at a very slow rate and in the two years of our own observations did not exceed one or two hundred feet. That there has been distinct forward movement and that it was in progress at the times of our visits was certain. In 1910 willows and alders up to 8 years of age were being overwhelmed by the advance of the ice margin. In June, 1911, the continuation of forward movement during the previous winter was shown by a series of snow arches, pushed up parallel to the edge of the glacier, and by the fact that,

where the snow had melted away, the advancing ice was overriding young shrubs which had budded in the season of 1911. There seemed to be no tendency to break up the buried ice in the interior flat.

At the glacier margin a low terminal moraine was being pushed along in front of the glacier, the slope of which at the end is 30° to 35°. That the forward movement was exceedingly slow was shown by the fact that this terminal moraine did not have the shape of a push moraine, as at Columbia Glacier, but resembled a talus. This was because a very small amount of outwash gravel was being ploughed up by the slowly-advancing ice compared with the larger amount of moraine which was sliding down from the surface of the glacier. The ice front was moderately crevassed, but the ice was rather dirty, basal ice with much detritus, which was being released by melting so that it was continually sliding down the ice front to the talus moraine. In places numerous ice blocks from crevassed pinnacles had rolled down the frontal slope of the glacier and lay on and in front of the terminal moraine.

Relation of Glacier to Railway. The advance of Grinnell Glacier, in 1910–1911, and possibly1912, might conceivably block the Copper River and Northwestern Railway, if the movement should continue long enough to break up the stagnant outer ice mass of the forested terminal hill and cause it to become crevassed, even if the glacier did not advance further into the river. This would make it impossible to operate the upper 150 miles of the railroad, for the small Grinnell Glacier, located at the lower end of Abercrombie Rapids, holds the key to the railroad situation from this point northward, particularly as the eastern bank of the Copper River in this portion (Pl. CLXV) is also a glacier, over which the building of a railroad would be very difficult and would involve at least two expensive bridges. The Grinnell Glacier is small, however, and the interior flat of moderate size, so that it would take a rather strong advance to disturb conditions in the outer part of the glacier where the railway crosses it and where practically all the ice seems to have melted out from beneath the present railway grade, though present a score or so of feet from it.

CHAPTER XXII

ALLEN GLACIER AND OTHER ICE TONGUES OF THE COPPER RIVER CANYON

The Canyon and Its Glaciers

The canyon by which the Copper River crosses the Chugach Range has a length of 100 miles. Two-thirds of this is north of Allen Glacier. In this portion its width varies from one to three miles. The valley walls rise to heights of 5000 to 7000 feet. Two large rivers join the Copper about half way through this canyon section, Bremner River coming in from the east and Tasnuna River from the west. These tributary valleys are of greater width than the trunk valley, but Copper River is of much greater volume than either of its affluents. In the Copper River valley south of the Bremner and Tasnuna rivers are Allen, Heney and several smaller glaciers. North of them the valley walls have a number of much smaller ice tongues. Allen Glacier, the first to be described in this chapter (Fig. 68), is immediately north of and close to Miles Glacier (Map 9, in pocket).

Allen Glacier

Name of Glacier. No name was given this glacier by any of the Russian explorers or by Allen, who recognized it as an ice tongue in 1885, but who named the portion of Copper River directly east of it Baird Canyon.[1] Hayes applied no name to the glacier in 1891, nor did Abercrombie in 1898, and it does not appear on the 1898 map by Mahlo. In 1900 Witherspoon applied the same name to it as the canyon,[2] and we have used his map and this name in our preliminary description.[3] There are three other Baird Glaciers in Alaska, however, one on White Pass, named by Schwatka in 1883 two years before Allen named Baird Canyon, the second in southeastern Alaska on Frederick Sound, named by Thomas in 1887, and the third east of Valdez, named by Schrader in 1898. A new name has, therefore, been declared necessary by the U. S. Geographic Board. It is proposed to use Allen,[4] after the intrepid explorer who was one of the first to visit it, at the time of his wonderful journey across Alaska in 1885.

General Description. Allen Glacier comes from unexplored snowfields in the Chugach Mountains west of Copper River, having a known length of over 5 miles, although undoubtedly it is longer, probably at least 15 or 20 miles in all. Its valley portion is a mile and three quarters wide. It has clear ice in its valley and there are few moraines on its surface. It flows eastward into the Copper River valley, expanding into a bulb five miles

[1] Expedition to the Copper, Tanana, and Koyukuk Rivers, Washington, 1887, Map 2; Narratives of Explorations in Alaska, Washington, 1900, map facing p. 434.

[2] House Doc. 546, 56th Congress, 2nd Session, 1900, Pl. II; Bull. 374, U. S. Geol. Survey, 1909, Pl. I.

[3] Tarr, R. S. and Martin, Lawrence, Nat. Geog. Mag., Vol. XXI, 1910, map on p. 25.

[4] This is Major Henry T. Allen, 8th Cavalry, United States Army.

wide in a north-south direction (Pl. CLXXIV), and spreading eastward completely across the valley, a distance of over three and a half miles. By this expanded glacier bulb Copper River is forced over against the eastern wall of the valley, just as it is forced to the western wall immediately below by the ice bulb of Miles Glacier. Opposite the end of the expanded glacier the stream is narrowed (Baird Canyon) and the current quickened, although there are no true rapids, and steamboats navigated this part of Copper River in 1909. The central part of the bulb is made up of clear ice and is lower than the eastern

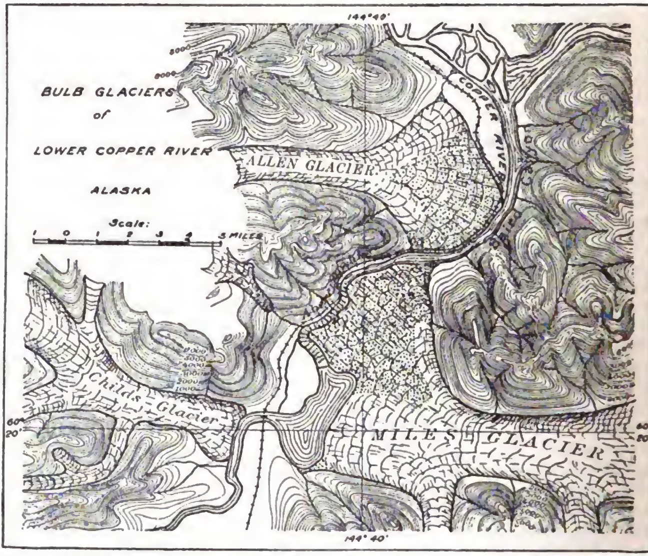


FIG. 68. ALLEN GLACIER AND ITS RELATIONSHIPS TO THE ADJACENT ICE TONGUES.
(After D. C. Witherspoon, U. S. Geol. Survey; the map first made in 1900 and republished with modifications in 1906.)

part, which is moraine-covered and clothed with vegetation. Copper River sweeps around the entire eastern and southern peripheries of the glacier, but on the northern side there is a large alluvial fan and a smaller one on the south side, separating the river and glacier. A rocky hill 600 feet high, rises out of the northern fan, and above it the river is broadened into a lake-like expanse.

Observations Previous to 1909. The Russians do not seem to have recognized this as a glacier, unless Grewingk also alludes to it as the "ice covered with earth," thought by us to refer to Miles Glacier. Abercrombie does not mention seeing the glacier in his 1884 exploration.

Allen does not show the glacier on his 1885 map upon which he named Baird Canyon, but says in the text,[1] that on April 4, 1885, "the camp at Baird's Canyon was at the foot of a vegetation-covered glacier which extended along the river for 6 or 7 miles. A short distance above the canyon the width of the river is two miles, with two small streams emptying into it on the east side. This widened part or lake extends about 6 miles."

In August, 1891, Hayes made the sketch map showing Allen Glacier, with "wooded moraine" on the terminus and the "shallow" lake above with "mud flats"; but he does not mention it in the text.

In October, 1898, Abercrombie speaks of the Copper River winding "around the glacial obstruction" of the "larger unnamed glacier," below which is Miles Glacier "likewise covered" with moraine and vegetation.[2] Guide Rafferty[3] in July, 1898, says "At the upper end of Baird's Canyon is a good-sized terminal moraine, back of which I supposed was a "dying" glacier. Baird's Canyon was found to be a very smooth piece of water, the river narrowing down from the wide bay to a few hundred feet in width."

Witherspoon's map, made in October, 1900[4] (Pl. CLXXIX) shows Allen Glacier much as now except that the eastern, moraine-covered portion of the bulb is not shown as part of the glacier but as a 600 foot hill between two forks of the glacier, which is made to bifurcate where it emerges from its mountain valley. This is surely not a change in condition but an error in observation, due to failure to recognize an alder-bearing, moraine-covered glacier.

In summary of the observations from 1885 to 1900, there seems to be no essential change; nor had there been any up to the time when the railway surveys of 1908 and 1909 discovered ice all along their 5½ miles of railway grade on the glacier. The bulb of Allen Glacier seems, therefore, to have been stagnant, moraine-veneered, and vegetation-covered throughout the period of visits by white men.

Our own studies of Allen Glacier were made by both of the authors for half a day in August, 1909,[5] and by the junior author[6] between August 24 and 27, 1910, when the contour map of Allen Glacier was made (Map 9). We also saw Allen Glacier in passing on July 4, 1911.

Valley Glacier. Allen Glacier flows from snowfields (Pl. CLXX) in the portion of the Chugach Mountains west of the Copper River and north of Mt. Williams[7] and other peaks at least 6000 to 7000 feet high. Within its mountain valley the glacier is a clean, moderately-crevassed ice tongue with a grade of 500 feet to the mile. On the northern side is a small lateral moraine and there is a much larger one extended along the southern

[1] Allen, H. T., Expedition to the Copper, Tanana, and Koyukuk Rivers, Washington, 1887, pp. 43–44; Narratives of Explorations in Alaska, Washington, 1900, pp. 425–6.

[2] Abercrombie, W. R., Reports of Explorations in the Territory of Alaska, 1898, War Dept., Adj.-Gen. Office, No. XXV, Washington, 1899, p. 320; also in Narratives of Explorations in Alaska, Washington, 1900.

[3] Rafferty, J. J., Same, p. 449.

[4] House Doc. 546, 56th Congress, 2nd Session 1910, Pl. I; also Bull. 374, U. S. Geol. Survey, 1909, Pl. I.

[5] Tarr, R. S. and Martin, Lawrence, Nat. Geog. Mag., Vol. XXI, 1910, pp. 14, 25, 27–28.

[6] Martin, Lawrence, The National Geographic Society Researches in Alaska, Nat. Geog. Mag., Vol. XXII, 1911, p. 541; Mastering the Alaskan Glacier Barriers, Scientific American Supplement, Vol. LXXI, 1911, pp. 305–307; Gletscheruntersuchungen längs der Küste von Alaska, Petermanns Geog. Mitteilungen, Jahrgang 1912, Augustheft, p. 79, Tafeln 9 and 11; Un Chemin de Fer sur Glacier dans l'Alaska, La Nature, Vol. 41, 1913, pp. 404–407.

[7] Named in 1910 for Mr. Alfred Williams, assistant engineer of the Copper River and Northwestern Railway.

margin. A medial moraine near the south side is broad and on passing outside the mountain valley it swings across into the middle of the glacier with a shape like a fishhook. At the mouth of the mountain valley the glacier grade is steepened slightly and the ice is more crevassed than in the piedmont bulb. This may be interpreted as evidence that Allen Glacier is in a hanging valley.

The Allen Glacier Bulb. While within its mountain valley the glacier has a width of only a mile and three quarters to two miles, it expands in the broad Copper River valley forming a bulb five miles wide (Pl. CLXXIV), with a grade which flattens to 180 feet to the mile. The medial moraine and the southern lateral, spreading out in this piedmont bulb, cover the southern part of the glacier with a veneer of moraine; but the northern part of the bulb, for a width of a mile and a half consists of almost-uncrevassed ice with practically no moraine covering. Here a few ice cones, with protecting moraine veneer, stand above the general level, as they also do near the center of the glacier at the margin of the clear ice area. The northern margin of the bulb has several medial moraines and a hummocky lateral moraine. These are all made up of black shale, greenstone, and graywacke, except one of yellow granite, and their prevailing dark color forms a striking contrast with the red moraines of the southern border. On the north much less of the margin of the bulb is moraine-covered. On each margin of the bulb close to the debouchure of the valley glacier there is a narrow barren zone of naked rock.

The Terminal Hill. East of this clear ice area, and rising to a height of a little less than 500 feet, is a wooded hill zone, which extends from the zone of barren ablation moraine about a mile eastward to the river. This terminal hill is the extension of the glacier bulb; although it does not look like a glacier, it is everywhere underlain by ice. Its highest part is nearest the clear ice, from which it is separated by an abrupt crescentic escarpment facing the glacier. The undulating eastern slope is a long gentle descent to the edge of the marginal channel of Copper River in Baird Canyon. This surface is rather hummocky and is slumping here and there, so that sections exposing bare ice, beneath ablation moraine and beneath a tangle of roots, are frequently revealed. Ice was seen in several places along the river bank in August, 1909, and many other exposures were seen when the railway was being constructed.

There are valleys cut in the moraine-covered glacier margin all along its outer periphery, in one case with a considerable basin holding a lake and doubtless due more to melting and slumping than to stream cutting.

Upon this stagnant, moraine-covered outer portion of the bulb grows a dense, continuous, mature, alder thicket, with scattered cottonwoods near the river. The annual rings in several of the largest alders were counted in 1910 and the ages varied from 50 to 67 years, indicating that the advance of the glacier across the valley, which we know from Allen's observations to have been over 25 years ago was certainly over 67 years ago (before 1843) and perhaps even earlier.

The Interior Flat. The clear ice area within the vegetation-covered bulb is an interior flat (Pl. CLXXII), intermediate in stage of development between that of Atrevida Glacier in 1909 and those of Variegated and Galiano Glaciers in 1905–09. In August, 1909, this interior flat was bounded by the crescentic escarpment of the forested terminal hill and by a terminal moraine. It represented the site of 200 feet or more of melting of clean ice, lowering below the former level, as indicated by the forested part of the ice

bulb. At the time of our observations in 1909 and 1910 there was an area of nearly-flat ice, merging into an area of moderately-rising glacier which sloped gradually upwards to the valley mouth. Upon the edge of this flat ice a broad stream was slowly flowing through a series of lake-like expanses in which clay and sand were being deposited upon the ice, as they had previously been deposited upon a similar but higher area to the northeast where slumping had followed the melting. A continuation of such alluviation upon the ice will produce a gravel-bottomed interior flat, similar to that of Variegated Glacier in Yakutat Bay. The ice surface in the flat, which was diversified only by broad swells and sags, had melted down to the bottoms of the crevasses. There was an occasional one in 1910 which had sloping walls and was 4 or 5 feet deep. Farther up the flat toward the valley glacier the crevasses were deeper but 30 feet was the maximum depth. Some of the shallower crevasses were filled with sand and mud. There were no appreciable changes in the ice or the drainage of the interior flat between 1909 and 1910.

The Northern Terminal Moraine. North of the bulb of Allen Glacier is a strong crescentic terminal moraine (Pl. CLXXII). It is over 3 miles long, an eighth to a quarter mile wide, and 50 feet high. Parts of it form a single broad ridge, other parts have shallow basins and flat knobs. It is separated from the glacier by a crescentic lowland a quarter to a half mile wide with pools, marginal channels, outwash deposits and lake clays, and slumping surface where these deposits had been laid down upon the nose of the glacier. This zone is barren of all vegetation.

The moraine seems to have no ice beneath it. Its surface is covered with moss and grasses and had alders and willows which were 9 to 19 years of age in 1910. One willow at the outer border of the moraine on the alluvial fan was approximately 50 years old.

The only gaps in this northern moraine were where streams from the glacier had cut through and only one of these was occupied in 1910. The other gaps were abandoned stream channels which headed up in the moraine or cut completely through it and terminated at a higher level than the present surface of the glacier.

Alluvial Fans of Outwash Gravel. The excess in the amount of thinning of the glacier by ablation in the unprotected interior flat has given the northern side a lower position than the moraine-protected southern margin and the main drainage from Allen Glacier, therefore, escapes this way, building the great alluvial fan, many times larger than that of the stream from the southern side. This northern fan (Pl. CLXXIV) slopes 54 feet to the mile in one portion and elsewhere 80 feet to the mile. It partly dams Copper River, aiding the glacier bulb in producing the lake-like expanse which backs up into side valleys and produces mud flats at low water and broad shallow expanses at times of flood (Pl. CLXXIX). The size of the alluvial fan on the north furnishes additional proof that the glacier has been where it now is for a long period; and the absence of vegetation on all parts of the fan shows that the streams have shifted back and forth over it regularly in building it up.

The Marginal Channel. The marginal channel of Baird Canyon (Pl. CLXXI) differs from that at Abercrombie Rapids in the lack of swift, white water with reefs and dangers to navigation, partly because the river is here in a broader channel and partly because the lake-like expanse above Abercrombie Rapids backs up into Baird Canyon and decreases the velocity there. On this account, and also because of the protection afforded by the large alluvial fan on the northern side of the glacier the river is not

making severe inroads on the glacier, as it does upon the Childs and the southern portion of Miles Glacier. No icebergs are discharged, for the glacier front is everywhere mantled by moraine and vegetation. In spite of the narrowing of the river from 8400 feet just north of Allen Glacier to 400 feet in Baird Canyon the current is not quickened sufficiently to make the river undercut the glacier front, although the ice is known to extend out to the river itself.

Evidence of Flowage in Allen Glacier. The layers of ice in many portions of the bulb of Allen Glacier are not horizontal, or inclined upward at a low angle, as near the termini of many valley glaciers. They are complexly folded and crumpled and, at many points within the interior flat, the general dip is vertical. This supports the idea of uprising ice currents within interior flats. The folds in several places (Pl. CLXXVII) were found to be thickened on crests and thinned on the limbs as in the type of what Van Hise has called *similar* folds.[1] As rock folds of this type are taken as evidence of rock flowage it seems possible that these ice folds may be interpreted as evidence of glacier movement by actual viscous flowage. The faults which cut certain folds are later than the folds and seem to have been made in the zone of fracture, or crevassing, after ablation of the glacier surface removed the overlying ice so that the folded ice was in the zone of fracture rather than in the zone of flowage. The ice is largely made up of coarse crystals.

History of Allen Glacier. The early stage of Allen Glacier as a tributary of the main Copper River Glacier seems to have been followed by a stage of complete deglaciation of the main valley so that when Allen Glacier expanded to its present form it had an open valley in which to spread out. This is suggested by the symmetrical mushroom head of this bulb glacier, which is the most perfect illustration of this type of glacier thus far described.

The date of this period of expansion is not known except that it was before 1843, as the vegetation proves. The time since this expansion seems to have been one of such slow movement in the valley glacier that the ice brought forward from the snowfields is practically all removed by vertical ablation in the interior flat. That the glacier is moving forward and is not "dead," as is commonly believed in Alaska, is necessitated by the snow supply in the mountains and proved by the crevassing in the portion of the interior flat adjacent to the valley glacier. The movement is very slow and does not seem to appreciably affect the outer, or eastern, portion of the bulb at all, for the vegetation has grown in the 4 or 5 feet of ablation moraine soil for more than half a century.

When we visited the Allen Glacier in 1909, bare ice was still visible in six of the railway cuts on the glacier terminus. A year later none of these exposures showed ice. The level railway grade, which had been produced in 1909 by blasting out a shelf in the glacier ice and levelling it up with morainic ballast, was seriously modified in 1910. The slumping of the grade, due to melting of the ice beneath, amounted to 2 or 3 feet in many places and, at the maximum, amounted to over 7 feet. Slumping at the water's edge showed that in 1910 the ice still extended out beneath Copper River. The largest area of slumping beneath the railway grade was 50 by 120 feet. Slumping had commenced here in 1909 and ice was then visible beneath the rails. The settling at this point amounted to 6 feet at one time of observation in 1910, although there had been some filling in the meantime by the railway section gang.

[1] Van Hise, C. R., Principles of North American Pre-Cambrian Geology, 16th Ann. Rept., U. S. Geol. Survey, Part I, 1896, p. 600.

The front of Allen Glacier has a distinct crenulation which is interpreted not as due to lobation of the ice front, but to slumping of areas where the ice melts out. One of these has already been spoken of as containing a good-sized pond. Another valley 30 or 40 feet deep was formed between our visits in 1909 and 1910 and in the latter year its margins showed long cliffs of bare black ice and a floor 100 by 300 yards strewn with dead alder. Earlier events of this sort were indicated by deep valley-like crenulations in the glacier front, some of them with mature alder growing on the sides where slumping had ceased, others with the slopes still covered with dead or dying vegetation.

Relation of Allen Glacier to the Railway. Nowhere else in the world, so far as we know, is a railway built for 5½ miles upon the end of a living glacier (Pls. CLXXI, CLXXXI, B). The ballast beneath the ties and rails of the railway actually rests upon the ice, not upon an abandoned moraine as at Heney Glacier. Since the railway was built, disadvantages in this location have been revealed, notably the settling of the track as the ice beneath it melts, the frequent breaking out of new streams, calling for new trestles, and in one case the shifting of a bridge support 18 inches toward the river, necessitating the driving of a new pile. These difficulties will recur every summer and will always render this section of the railway expensive to maintain and slow for trains to run over. Of course there is no danger to passengers on a railway thus located, provided track-walkers watch the railway grade carefully.

The gravest problem in connection with the section of the railway on Allen Glacier comes in connection with a possible advance of the glacier. This would destroy the track and completely block traffic on the whole railway, for there is no way to go around. After a period of immunity of 67 years such an advance is likely to come at any time and may be especially imminent any time within the next few years,[1] for the adjacent Childs, Grinnell, and Heney Glaciers had advances in 1910 and 1911, suggesting that the snow-fields west of Copper River are likely to also produce an advance in the Allen Glacier, which lies between the Grinnell and Heney ice streams.

Such advance, however, would probably be short-lived and, if not too large, its impetus might all be taken up in the interior flat, so that the outer edge of the glacier bulb, on which the railway is built, need not necessarily be broken up.

The railway engineers could not have avoided this stretch of track upon the glacier, without prohibitive expense of rock cuts, a tunnel, and two expensive bridges across Copper River. The most serious error in the railway building was the stripping off of all vegetation from the right of way. If the alders had been left growing close up to the railway the slumping of the ice beneath and close to the track would have been much less. Even now it would probably pay to plant the right of way with new shrubs so that further slumping beside the track may be decreased. It seems likely that renewal of ballast beneath the rails will eventually bury the ice there so deeply that slumping will cease.

[1] Since making this prediction of imminent advance of Allen Glacier the junior author has been informed by Mr. Caleb Corser of the Copper River and Northwestern Railway that during the summer of 1912 the glacier commenced to move forward. The southern border was crevassed and broken in September, 1912, and showed clear blue ice where there was smooth, moraine-veneered ice in 1911. In the northern, moraine-covered margin of the bulb there was much thickening and crevassing and an advance which is said to amount to ⅓ mile. The largest glacial stream on the northern side of Allen Glacier left the channel which it occupied from 1909 to 1911 and flowed in a new channel, in 1912, about a mile farther west. Its volume was greatly increased and the stream shifted frequently, causing much trouble in the portion of the railway grade on the northern alluvial fan of outwash gravels.

Small Ice Tongues North of Allen Glacier

La Gorce Glacier. In the valley immediately east of the terminus of Allen Glacier there is a small ice stream called La Gorce Glacier.[1] It is a valley glacier a little smaller than Grinnell Glacier and terminates about 2 miles from Copper River. This ice stream flows with a rather steep grade from a large cirque. Its terminus is mantled with débris. There is a strong lateral moraine on the valley side east of the glacier margin, which seems to be of recent formation. A broad valley train of outwash gravels leads from the glacier to the Copper River.

The mouth of the valley of La Gorce glacier is exceptionally steep-sided because of glacial erosion. It seems clear that the La Gorce valley formerly mouthed at least a mile farther north and that the Copper River now sweeps through the mouth of La Gorce valley (Pl. CLXXIV) in the portion of its course south of the 600 foot rock hill near Allen Glacier. This hill was formerly the termination of a spur on the northern side of La Gorce valley. Allen Glacier seems to have forced the Copper River into a marginal channel which crossed this spur and entered La Gorce valley, the river then sawing off the end of the spur and converting it into a detached rock. It is concluded that this diversion was related to an earlier expansion of Allen Glacier than the present one because the channel east of the rock hill is not a narrow, steep-sided gorge, but has been opened out to a width of a quarter mile with flaring walls. This is ascribed to erosion by the former trunk glacier of the Copper River valley. The expansion of Allen Glacier is thought to antedate this, and the reoccupation by Copper River of this modified marginal channel may be due to alluviation by the glacial streams on the northern side of Allen Glacier.

Wernicke Glacier.[2] Wernicke River enters the Copper River valley northeast of Allen Glacier, flowing in scores of channels over a gravelly valley train a mile wide. We have not visited the glacier at the head of this valley but from the many streams it is judged to be of considerable size.

Shields Glacier. Shields Glacier [2] is a small valley glacier on the western side of Copper River valley about 6 miles north of Allen Glacier. It terminates near the lip of its hanging valley (Pl. CLXXIX).

Smaller Ice Masses. A hanging valley south of Shields Glacier still retains a small glacier. There are several on the valley wall north of the La Gorce Glacier, two of which come so nearly in line that they give the erroneous impression of a recemented glacier.

North of Wernicke River there is a névé-sheathed slope of the type seen in Harriman Fiord. This ice expands on a moderately steep slope and terminated in 1910 just above a barren zone, where there was formerly a steep ice cascade.

Heney Glacier

Location. Heney Glacier is on the western side of Copper River 10 miles north of Allen Glacier. The portion of the Copper River canyon near Heney Glacier is 2 to 4 miles wide with the mountain walls on each side rising precipitously over 5000 feet. Instead of being blocked by glaciers that project clear across the valley from either side, as Allen

[1] Named in 1910 for John O. LaGorce, assistant editor of the National Geographic Magazine.
[2] Named in 1910 for L. Wernicke and Archie Shields of the Copper River and Northern Railway.

and Miles Glaciers do, the canyon is open, having only this one large glacier extending into it, and this ice tongue reaches only about a mile into the Copper River valley. The other glaciers of this portion of Copper River end at a considerable height in the mountains.

General Description. This ice tongue rises in cirques and snowfields 4000 to 6000 feet above sea level. It has a known length of about 8 miles and perhaps an even greater extension in the mountains. At the mouth of its mountain valley it has a width of about a mile. Outside the mountain valley it expands in a bulb over two miles wide, and originally similar in form to the bulb of Allen Glacier. The bulb of Heney Glacier has now been modified by retreat, but more is left of it than that of Childs Glacier, because Copper River has not cut into the Heney and modified its form.

McCune Glacier, a former tributary of Heney Glacier on the south side, erroneously shown on previous maps as confluent with Heney Glacier, has been disconnected from the main ice tongue by melting and now terminates about half a mile within its valley.

Heney Glacier terminates in a low slope, most of which is covered with ablation moraine and supports thick vegetation. In front of the glacier is an abandoned terminal moraine, traversed in a great curve by the railway. Between the railway and the glacier there are several lakes, the shores of two of which are traversed by the railway. In the largest one, which is farthest to the north, the glacier has a vertical cliff and discharges a few small icebergs. Heney Glacier is nearly as large as Childs Glacier, but is much less impressive as seen from the railway, because it is not so active, and has not a high vertical cliff undercut by the Copper River. Because it is moving slowly, the outer end of it is dark-colored and unattractive, with ablation moraine and ice covered with thickets of alder and cottonwood. The upper part of the glacier is clean and beautiful and a view from a point about a mile west of the railway shows Heney Glacier as an impressive and attractive tongue of ice, ranking fully with the Childs Glacier in size and beauty.

Previous Maps. Heney Glacier has not been described previously. It was first shown upon a map by Mahlo in 1898.[1] It was sketched more accurately by Witherspoon in 1900 [2] (Pl. CLXXIX) and was shown upon a later map of the U. S. Geological Survey.[3] The detailed map of the terminal bulb (Pl. CLXXVIII) was made in 1910 by the National Geographic Society's expedition. Rough sketches of the relationship of the glacier terminus were made by the railway surveyors between 1906 and 1909, and their maps are the first ones upon which the presence of the lakes in front of Heney Glacier is shown. Throughout this period of observation, this ice tongue seems to have remained in about the same position and condition, and it may be assumed to have been inactive, with normal slow motion in the valley glacier and tributaries, and with an almost stagnant condition in the bulb, from sometime before 1898 to the summer of 1910 when our studies were made.

The Valley Glacier. The valley glacier showed no abnormal features in 1910. It heads on a snow divide with the Schwan Glacier of Tasnuna valley, flowing eastward and receiving as tributaries two cascading glaciers from the west and two large and two small glaciers from the south. The slope of the main ice tongue is moderate, an ascent

[1] Abercrombie, W. R., War Dep't., Adj.-Gen. Office, No. XXV, Washington, 1899; Schrader, F. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, Map No. 21.

[2] House Doc. 546, 56th Congress, 2nd Session, Washington, 1900, Pl. II.

[3] Bull. 374, U. S. Geol. Survey, 1909, Pl. I.

from 580 feet to 830 feet above the railway being made in the two miles from the mouth of the valley glacier to the end of the first large tributary on the south. The surface of the valley glacier is little crevassed and only slightly encumbered with moraine. The moraine extends farthest up the glacier on the northern margin. The crevasses were predominantly longitudinal and travel was easy in 1910 upon the long narrow seracs of the valley glacier.

Barren Portion of Bulb. The southern half of the bulb of Heney Glacier is free from moraine and slopes 10° to 20°. There is a little débris in this basal ice, but the surface has scattered stones upon it. Near the glacier margin the bottoms of a few crevasses reach to the very bottom of the glacier, revealing ice layers which stand at a high angle.

The northwestern half of the bulb is thickly mantled with ablation moraine and its surface is a maze of knobs, kettles, ridges and hollows, with ice showing in little cliffs at a few localities. The surface rises gradually from 150 to 580 feet above the railway. The surface of this ablation moraine was everywhere slumping in 1910 and vegetation had nowhere taken root upon it, except in a narrow strip close to the forested portion of the bulb. Even here the plants were very young and were scattered, so that, in general, there was a sharp line of demarcation between the barren and the forested zones of the moraine-covered bulb.

Forested Portion of Bulb. All of the eastern end of the bulb (Pl. CLXXVIII) is thickly covered with ablation moraine, beneath which the presence of ice is only shown by infrequent areas of slumping. This portion of the glacier is completely covered with dense forest. Most of the trees are alders, but there are scattered cottonwoods and willows.

The Terminal Moraine. The terminal moraine (Pl. CLXXVIII) is followed by the railway from the lake south of Mile Post 75 to the curve just beyond Mile Post 77. The glacier has retreated half to seven-eighths of a mile from the moraine. Near Mile Post 75 and for over a mile to the south the glacier retains an advanced position so that no moraine has been exposed by melting. For a short distance beyond this, however, there is an independent morainic ridge, from which the glacier has receded about half a mile. This southern moraine is broad and low and partly covered with forest.

The northern portion of the moraine where followed by the railway, varies in width from a single, symmetrical, narrow ridge to a moraine belt nearly an eighth of a mile wide. The height ranges from 5 to 15 feet, for the lower ridges, to 40 or 50 feet. The ridges are crescentic and parallel to the former ice front, and sometimes contain small ponds and dry kettles. The material in different parts of the moraine is till, rounded outwash gravel, fine river silt, cross-bedded dune sand, and huge angular bowlders. In several localities the till overlies the dune sand or the large angular blocks rest upon stratified stream deposits, showing that there has been an advance after a period of retreat. In no case were the lower deposits at all weathered.

The Marginal Lakes. There are four large and several small lakes around the margin of Heney Glacier. These lakes have one wall of ice and one of terminal moraine. Three of them lie within the two areas of interior flat which this bulb glacier has developed upon either side of the central ridge of ablation moraine. That the lakes have fluctuated in level is well shown in the southernmost, where there were minute, abandoned beach levels in 1910.

The southern interior flat contained an area of ground moraine which was crossed by a remarkable series of hundreds of low, parallel furrows and ridges (Pl. CLXXX) of

Wooded Surface of the Western Part of the Stagnant Bulb of Miles Glacier
In the zone of thickest ablation moraine. From Station N, August 10, 1910.

Plate CLXII

Surface of Miles Glacier in the Zone of Thick Ablation Moraine

The steep slopes are glacier ice overlain by morainic débris with only a few, scattered small shrubs. Dense vegetation of zone of thickest moraine in background. Photograph, August 22, 1910.

PLATE CLXIII

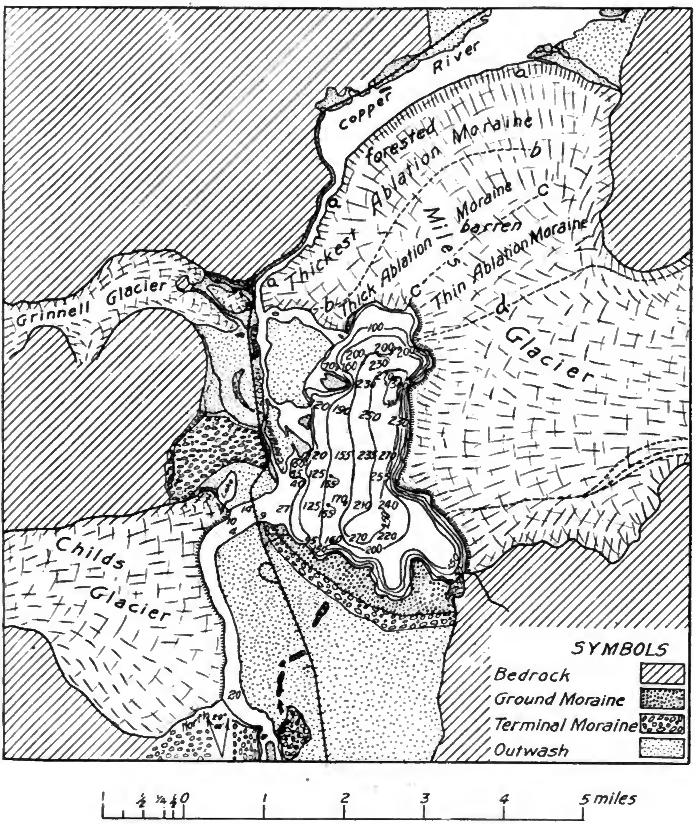
ZONE OF THIN ABLATION MORAINE ON BULB OF MILES GLACIER
Photograph, August 22, 1910.

Plate CLXIV

The Expanded Northern Bulb of Miles Glacier

Letters indicate boundaries of the divisions of the bulb: a–b, zone of thickest ablation muraine; b–c, zone of thick ablation moraine; c–d, zone of thin ablation moraine. Photograph from Station E, August 21, 1910.

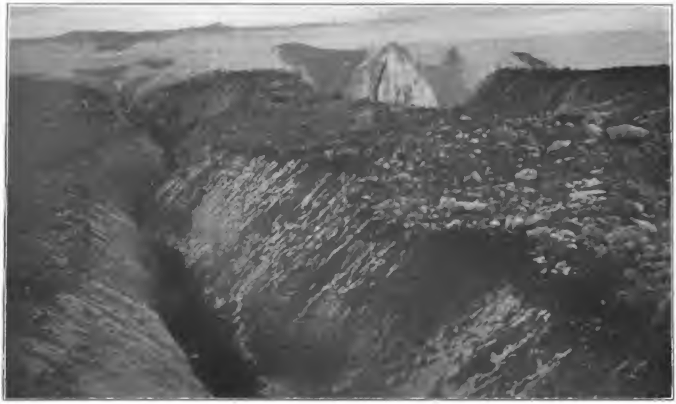
Plate CLXV



Miles and Childs Glacier Bulbs

Plate CLXVI

Ground Moraine Surface between Southern Terminal Moraine and Miles Lake
The angular, unstriated rocks in the foreground are interpreted as part of a long, narrow crevasse deposit. Photograph, August 22, 1910.



A. CREVASSE IN DETACHED ICE MASS AT SOUTHERN EDGE OF MILES GLACIER
With cover of ablation moraine sliding into it. Photograph, August 19, 1910.

B. THE FORESTED TERMINUS OF GRINNELL GLACIER IN 1910

PLATE CLXVIII

RETREAT OF MILES GLACIER FROM 1900 TO 1907

Followed by advance between 1908 and 1910. Photograph by F. C. Schrader, from Photo Station M, October 6, 1900.

PLATE CLXIX

TERMINUS OF CLEAN PORTION OF GRINNELL GLACIER FROM STATION O ON AUGUST 20, 1910

Plate CLXX

The Allen Glacier

Emerging from its mountain valley and spreading out in its piedmont bulb. Mt. Williams in the background. Photographed from Station P in 1910.

PLATE CLXXI

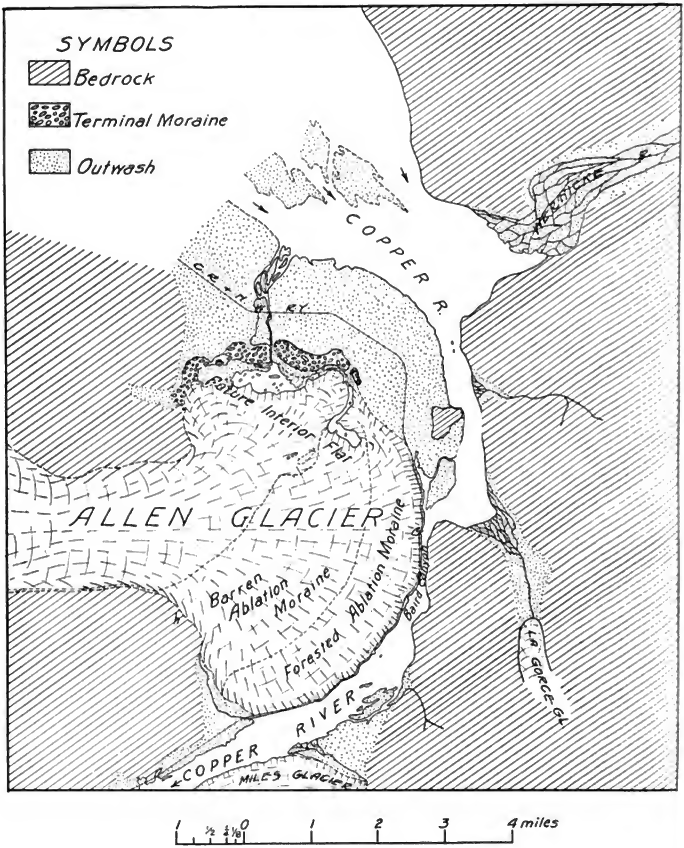
THE COPPER RIVER AND NORTHWESTERN RAILWAY
On the stagnant ice of Allen Glacier. Copper River on one side, ice beneath, glacial ice with moraine and forest on the right.

Plate CLXXII

The Northern Terminal Moraine and Part of Interior Flat of Allen Glacier in 1910 from Station P

A. Ridges in Southern Interior Flat of Heney Glacier
Terminal moraine in right background.

B. Hanging Valley North of Heney Glacier



ALLEN GLACIER IN 1910

PLATE CLXXV

THE UPPER NARROW PORTION OF THE COPPER RIVER CANYON ACROSS THE CHUGACH MOUNTAINS
Deeper in this portion than the Grand Canyon of the Colorado. Photograph by E. A. Hegg.

Plate CLXXVI

A. Copper River near Chitina
Looking south toward the Hanagita watergap in which Wood Canyon is cut.

B. Looking North in Wood Canyon
Hanagita watergap at a, post-glacial gorge of Copper River at b, intermediate glacial valley at c. Photograph by E. A. Hegg.

the type formed by a minor advance of a glacier across the unconsolidated deposits of a previous retreat. These parallel ridges were crossed at various angles by narrow deposits of sand and clay. Some of the latter were 3 or 4 feet high (Pl. CLXXIII, A) and rather sinuous. Those at the glacier margin had cores of ice.

The marginal lake at the eastern end of the glacier is the first one seen from the train going northward. Its ice wall is completely covered with growing forest and slopes at a moderate angle.

The northern marginal lake, the largest bordering Heney Glacier and also well seen from the railway, is about a mile long and half to three quarters of a mile wide. It has the terminal moraine on the north and east sides, outwash gravels on the west, and the glacier on the south. This glacier margin, which is in the zone of barren ablation moraine, has been undercut by the lake waters, so that it forms a precipitous cliff, 20 to 50 feet high. At the time of our studies ice blocks were occasionally sliding down into the water, more at night than during the day. The lake is probably very shallow, for no icebergs were afloat upon it.

The Outwash Gravels. The deposits made by the glacial streams from Heney Glacier merge with those of the Copper River. The largest alluvial fan is on the south and is barren near the present streams and thickly covered with alder elsewhere. The modern outwash gravels west of the largest lake are barren and surround remnants of older outwash which is forested.

Vegetation on Heney Glacier. The trees and moss on the portion of the glacier which is covered with ablation moraine and upon the terminal moraine furnish some facts regarding the recent history of the terminal bulb. The absence of annual plants and of moss on the western belt of ablation moraine show that melting of the ice has resulted in slumping for some time in the past. The presence of moss and of a dense growth of mature trees on the eastern extremity shows that it has long been immune from slumping, and yet, even here, the rare areas of dead and overturned trees show that the ice lies at no great distance below the surface. Upon the outer margin of the glacier and upon the terminal moraine near Mile Post 75 the alders and cottonwoods were 11 to 26 inches in diameter and 50 to 76 years old in 1910, but at the extreme northwestern edge of the terminal moraine, near Mile Post 77 there were cottonwoods over 90 years old. This shows clearly that the last expansion of the bulb of Heney Glacier was about a century ago and that for over 76 years there has been no period of activity capable of breaking up the outer portion of the bulb.

Near the outlet of the large northern lake the crest of a high knob in the terminal moraine showed great cracks in 1910. As the rocks were thickly covered with moss and the ridge was forested this raised the question as to whether the ice still underlies this moraine and the lake basin. We saw nothing elsewhere that led us to suspect ice beneath this moraine, the remainder of which is forested and undisturbed by slumping.

Activity of Heney Glacier in 1911. It has been stated that the glacier was only moderately crevassed in 1910. The crevasses in the valley tongue seemed to be of two sorts, one set were the normal crevasses of the more rapidly moving upper glacier reduced by ablation, the other a longitudinal series of fresh-looking gashes which had sliced the glacier into long flat-topped seracs. This latter series extended into the edge of the zone of barren ablation moraine, where none of the older crevasses were left. That

they were made in the season of 1910 was evident from their relationships to the ablation moraine. They died out within a few hundred feet of the edge of the clear ice.

In July, 1911, Heney Glacier was seen from the railway and the crevasses then extended throughout the zone of barren ablation moraine. They were much more abundant in the clear ice of the southern portion of the bulb and extended clear to the margin of the large northern lake. It was evident that portions of the glacier easily traversed in 1910 were nearly, if not quite, impassable in 1911. Mr. L. Wernicke, one of the railway engineers, noted in September, 1911, that there were many icebergs in the northern lake.

Evidently a wave of abnormal movement spread down through the Heney Glacier in 1910–1911. It seemed to be only a minor spasmodic advance of the sort which prevents vegetation from gaining a foothold in the inner portion of the bulb, rather than a great advance like the one which resulted in the building of the terminal moraine about a century ago.

Adjacent Glaciers. Of the ice tongues near Heney Glacier the largest is the McCune Glacier, already alluded to as a detached tributary of the Heney. This valley glacier is about an eighth of a mile wide and over 4 miles long, rising in snowfields adjacent to those of Shields Glacier. In 1910 it was moderately crevassed and had a strong medial moraine and two lateral moraines. The half mile between this glacier and the margin of the Heney is occupied partly by outwash gravels and partly by two independent moraines, one a terminal moraine of McCune Glacier, the other a lateral moraine of Heney Glacier, the margin of whose bulb has bulged into this valley since the McCune Glacier became independent.

There is a small glacier in the tributary valley north of Heney Glacier (h, Pl. CLXXIX). This glacier formerly supplied ice to the Heney and now hangs 600 or 800 feet above the surface of the trunk glacier. The water from the small glacier descends from the lip of the hanging valley in a series of beautiful waterfalls (Pl. CLXXIII, B).

The first cirque south of Heney Glacier contains a small stagnant ice mass, covered with ablation moraine. This cirque hangs over 1000 feet above the floor of the Copper River canyon.

Other Glaciers of the Copper River Canyon

The ice tongues of the Copper River canyon north of Heney Glacier are not at all well known. We have studied none of them, but have seen several of them from a distance. At the head of the Bremner River there are large glaciers, and on the south side of the Tasnuna valley there are two large bulb glaciers, of which Woodworth Glacier seems to be larger than Schwan Glacier. Each of the latter was stagnant and moraine-covered in 1910 and apparently much as they were when visited by Schrader [1] in 1898.

Most of the ice tongues north of the Tasnuna River, except Cleave Creek Glacier,[2] are small and descend at steep grades, in several cases as cascading glaciers, as on the peak south of Dewey Creek. The noteworthy feature about all these ice tongues is their small size in contrast with the great size of the glaciers farther south in the Chugach Mountains where the snowfall is many times greater.

[1] Schrader, F. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 364, 397–398, Pls. XXXII, XXXIII and Map 20.

[2] *Ibid.*, pp. 362, 396 and Pl. XXX.

CHAPTER XXIII

GLACIATION OF THE PRINCE WILLIAM SOUND AND LOWER COPPER RIVER REGIONS

Introductory

This chapter will be devoted to a discussion of some of the phenomena of general glaciation of the Prince William Sound and lower Copper River regions, except as already taken up in connection with individual glaciers and separate fiords. As in the corresponding discussion for the Yakutat Bay region (pp. 198–231), there will be amplification of special topics, but no attempt will be made to summarize the phenomena of glaciation presented in (pp. 232–450), except as they furnish evidence on the problems under consideration.[1]

The Copper River canyon will be dealt with first, then Prince William Sound, and, lastly, there will be a brief comparison of the glacial history of the Prince William Sound and Yakutat Bay regions.

The Lower Copper River

The Canyon Between Chitina and Tasnuna Rivers. From the broad intermontane basin of Copper and Susitna Rivers, which is surrounded by the Chugach, Wrangell, Talkeetna, and Alaska Ranges there are water-gaps and passes leading out in various directions. Copper River emerges at the lowest of these and plunges into the hundred miles of canyon by which this stream crosses the Chugach Coast Range to the Pacific Ocean. This canyon consists of three quite different portions, (1) a northern, narrow canyon, (2) a wider, middle canyon, and (3) a southern, terminal portion which flares open toward the ocean. The northern, narrow portion of this canyon lies between the Chitina and Tasnuna Rivers. Its topography is shown in Figs. 69, 70, and Pl. CLXXIX.

From the Chitina River southward (Fig. 69) the river valley at first narrows downstream (Pl. CLXXVI, A), the valley walls become higher, and the river plunges into Wood Canyon, a short, narrow, steep-walled gorge (Pl. CLXXVI, B).

Below Wood Canyon the river is in a broader, young, stream valley, whose walls rise still higher as the stream penetrates the mountains, until, near Spirit Mountain and Tiekel River, elevations of 6000 to 7000 feet are attained within 2 miles of the river. It is, thus, even deeper than the Grand Canyon of the Colorado River in Arizona, but differs from it in having a form produced by glacial erosion. With this general character (Pl. CLXXV) the canyon continues southward to Tasnuna River. Its grandeur is not seen as well from the railway, which follows the western bank, as from

[1] For general maps of the area discussed see A. H. Brooks, Pl. 5, House Doc. 1346, Part 2, 62nd Congress, 3rd session, Washington, 1913; U. S. Grant and D. F. Higgins, Pl. II, in pocket, Bull 443, U. S. Geol. Survey, 1910; and Map 1, in pocket of this book.

the river itself, and there are few finer river trips in the world than that from Chitina River southward through the Copper River canyon. The junior author had the opportunity of making this in 1910, when steamboats were running on the Copper.

The present grade of the river in this upper portion of the canyon ranges between 6 and 8 feet to the mile. The river is an anastomosing stream, divided into scores of shallow, braided channels on a narrow valley bottom of glacial outwash gravels. The only exception to this is the short stretch of Wood Canyon, where the river covers the whole bottom of a single, deep gorge. Copper River is heavily laden with glacial débris from the Wrangell, Chugach, and Alaska Ranges and the bottom of the canyon is being built up, so that the channels are continually shifting. One day in August, 1910, for example, a steamboat upon which the junior author ascended and descended the Copper River for 30 miles was obliged, on the return trip, to seek new channels at several points where deposition had shallowed the channels which were navigable earlier the same day.

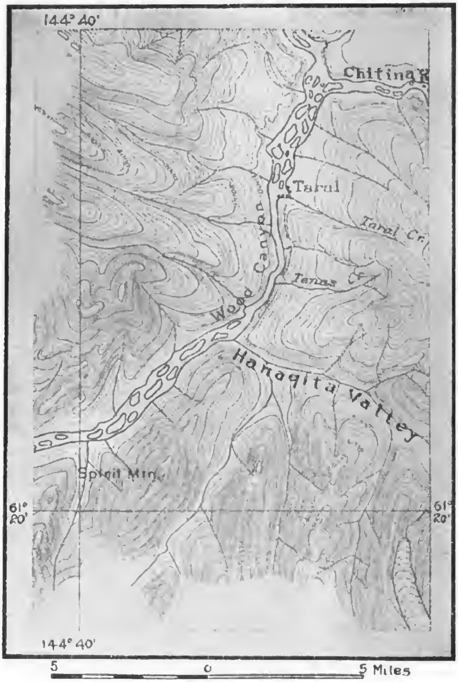


Fig. 69. Copper River Canyon.
(After U. S. Geological Survey.)

The portion of the Copper River canyon between Chitina and Tasnuna River has been intensely glaciated. The walls are much oversteepened (Fig. 70) and the U-shape (Pl. CLXXIX) is well developed. The oversteepening results in modern avalanches, as in a case just south of Cleave Creek. There are numerous hanging valleys (see h, Fig. 70), which are at elevations of from 300 to 400 feet for the larger tributaries, increasing to an extreme of 1500 to 2000 feet in the case of a small hanging valley just north of Dewey Creek. The post-glacial gorges cut in the lips of these hanging valleys

contain numerous beautiful waterfalls. Hanging valleys of the second order are also present. The hackly surfaces characteristic of glacial plucking are prominent on some of the oversteepened slopes. The roches moutonnée form is best developed in the hill of naked rock at the mouth of the Tiekel River where the east-west movement of this large glacial tributary has predominated. The upper limit of glaciation must be 3000 to 5000 feet in the northern portion of the canyon. Above this the irregular superglacial surfaces seem to be due more to weathering during the extreme glaciation than at present, for there are no great accumulations of talus below these precipitous cliffs, as would be the case if post-glacial, high-altitude weathering had produced them, rather than super-glacial weathering with a glacier present below in the canyon to transport the weathered material away.

The canyon winds more than in the broader section south of Tasnuna River, so that one sees overlapping spurs from certain points, but the ends of these spurs are sharply truncated by glacial erosion. The cross-section of the gorge is in some places asymetrical, as in the eight miles from Tiekel to Cleave Creek, where the eastern side of the canyon is much more oversteepened than the western side, which still retains some long, sloping, overlapping and only slightly truncated spurs.

Fig. 70. Copper River Canyon.
(After U. S. Geological Survey.)

The same section also shows the rock terrace in which the present stream course is cut in some places, and which, here and there, produces a system of very low, overlapping spurs.

For about fifteen miles south of Wood Canyon Copper River occupies an ill-defined secondary gorge within its glaciated canyon. The gorge is cut in a rock bench or ter-

race which slopes gradually southward from the mouth of Hanagita Valley. Young gorges have been cut across the remnants of this bench by tributary streams. Its top has the rounded outlines of a glaciated surface. Near the southern portion glacial gravels rest upon it.

The isolated 2200 foot rock hill called The Peninsula (Pl. CLXXIX), and a similar hillock just south of Cleave Creek, present interesting problems, deserving of attention, in connection with former marginal drainage and glacier distribution.

The Origin of Wood Canyon. Wood Canyon is a young, steep-sided gorge of a sort seen nowhere else in the lower Copper River. Its origin is interesting and may possibly be of importance in connection with the problem of how Copper River acquired its present course across the Chugach Range. This problem has been touched upon by Schrader and Spencer,[1] and by Mendenhall.[2]

Without going in detail into this larger problem, which requires further field work in relation to several other outlets of the Copper-Susitna basin, it may be pointed out that a glacial origin[3] seems more reasonable than headwater erosion, as suggested by Mendenhall, or an antecedent origin for the course of Copper River through this northern portion of the mountains, with the Wood Canyon interpreted, as has been done by Schrader and Spencer, as an evidence of recent warping.

This hypothesis of glacial origin of the course of Copper River across the mountain range may be outlined as follows: The preglacial upper Copper and Chitina Rivers are thought of as having some other course to the sea than the present one. The preglacial lower Copper River is thought of as heading somewhere south of but close to the present Wood Canyon. A former subsequent stream in the Hanagita Valley is thought of as turning at right angles at the south end of what is now Wood Canyon and flowing northward through the pass, alluded to hereafter as Hanagita Watergap, to Chitina River, just as Tebay River fifteen miles farther east, now flows northward from Hanagita Valley to Chitina River.[4] The hypothesis then supposes that the intermontane glacier of the Copper River basin rose high enough to have an outlet southward through the Chugach Mountains by way of the lower Copper River, and that a glacial stream cut the present Wood Canyon and diverted the drainage of the eastern portion of the Copper River basin, that is the Upper Copper and Chitina Rivers, into the headwaters of the lower Copper, whose valley had been enlarged by glacial erosion to the present canyon form.

The conditions at Wood Canyon are as follows. The canyon is a narrow gash (Pl. CLXXVI, A) cut in the bottom of an older col. This broad flat-bottomed col, Hanagita Watergap, seems to be a preglacial stream course at a little higher level than the broad rock terrace in which the present gorge of the adjacent Chitina River is in-

[1] Schrader, F. C., A Reconnaissance of a Part of Prince William Sound and the Copper River District, Alaska, in 1898, 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 395–403; Schrader, F. C. and Spencer, A. C., Geology and Mineral Resources of a Portion of the Copper River District, House Doc. 546, 56th Congress, 2nd Session, Washington, 1901, pp. 70–75; Spencer, A. C., Pacific Mountain System in British Columbia and Alaska, Bull. Geol. Soc. Amer., Vol. XIV, 1903, pp. 120–121, 127.

[2] Mendenhall, W. C., A Reconnaissance from Resurrection Bay to the Tanana River, Alaska, in 1898, 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, pp. 334–335.

[3] Martin, Lawrence, Nat. Geog. Mag., Vol. XXII, 1911, p. 560; Petermanns Geog. Mitteilungen, Jahrgang, 1912, p. 81; The Canyon and Delta of the Copper River in Alaska, Bull. Geol. Soc. Amer.: Vol. 24, 1913.

[4] See Pl. I., Bull. 374, U. S. Geol. Survey, 1909.

cised.[1] From this difference in level, it is thought that the preglacial drainage through the Hanagita Watergap went northward. The watergap is thought of as widened out to the U-shape by the ice which streamed southward from the Copper River basin.

The glacial erosion of the watergap produced a secondary valley within it, but this glacial valley nearly everywhere coincides with Wood Canyon and, therefore, is recognizable at only one point, where about half a mile of it is preserved. This is where the Railway leaves the west bank of the river, just north of Mile Post 124, and follows this abandoned valley (c in Pl. CLXXVI, B, and Fig. 71) which is 80 or 90 feet above the present level of the river.

FIG. 71. COPPER RIVER IN WOOD CANYON.

When the glacier was melting away this glacial valley was apparently occupied by a tongue of stagnant ice long enough to prevent the main stream from flowing here, for part of the Wood Canyon channel was established just to the east. That glacial waters occupied this channel at one stage, however, is shown by outwash gravels 90 feet above the present Copper River.

Wood Canyon (Pl. CLXXVI) was afterward incised to its present depth and is chiefly a postglacial gorge. In it the Copper River is constricted in a channel 400 to 550 feet wide, in contrast with 1300 feet just below the canyon and over half a mile just north of it.

The origin of Wood Canyon is clearly involved with the glacial period rather than orographic movements. The great terraces of outwash gravels in the Copper and Chitina valleys are graded up to an elevation that makes it impossible to consider their formation in relation to anything but streams graded up to a temporary baselevel at the height of the top of the Wood Canyon walls, or the approximate height of the abandoned valley followed by the railway. This establishes Wood Canyon as post-glacial, and suggests the control of the downcutting of the rock floor of the canyon over the system of terraces and gorges which have been cut in the glacial gravels of the whole Copper River basin.

How long ago Wood Canyon was eroded cannot be stated, nor is it likely that downcutting has entirely ceased. It seems certain, however, that it has taken several thousand

[1] See Pl. L, Bull. 374, U. S. Geol. Survey, 1909.

years, for the loess deposits overlying the outwash gravels immediately north of Wood Canyon have been at least 1000 years, and probably much longer, in process of accumulation.[1]

The Canyon Between Tasnuna River and Childs Glacier. The middle portion of Copper River Canyon extends from Tasnuna River to Childs Glacier (Pl. CLXXIX). The width in this section is from 2 to 4 miles, in contrast to the width of a half mile to a mile in the narrower canyon between Tasnuna River and Chitina. The width in this glacial canyon does not increase down-stream as in a valley carved by river erosion. Just below Tasnuna River, for example, the canyon is over four miles wide, and fourteen miles to the south at Allen Glacier it is four miles wide, while six miles farther down-stream the width is only two and one-half miles, and six miles below this, at Childs Glacier, the width is again four miles. The canyon walls are slightly lower than in the narrow, northern section, the average height in this middle section being 5000 to 6000 feet, attained within 1 to 2 miles of the river. The U-shape is well developed although the canyon walls flare more than in the northern section.

The broadened valley bottom of the Copper River canyon was recognized by Allen in 1885 as an evidence of profound glacial erosion.[2]

The cause for the increased width of the canyon from Tasnuna River to Childs Glacier seems to be the incoming of the Tasnuna and Bremner valleys. The Tasnuna, about two miles wide, and the Bremner, about a mile and a half in width, join the Copper, which is less than a mile wide, causing the four miles of width in the section of the Copper under discussion. This matter of width is another argument in favor of the recency of the diversion of the drainage of the Copper River basin across the Chugach Mountains. The relationship of widths of the Copper and these two tributaries is a natural one, granting profound glacial erosion by the Copper River Glacier, aided by the large former ice tongue in the Tasnuna valley, which still contains the Woodworth and Schwan Glaciers, and the large, former ice tongue of the Bremner valley, where there are great glaciers even now.

The canyon walls between Tasnuna River and Childs Glacier are exceptionally simple, chiefly as a result of glacial erosion. There are no projecting spurs (Pl. CLXXIX), except the tiniest remnants of eroded spurs, as in the one in the rock cut between Allen Glacier and Camp 55 at Abercrombie Rapids. The lower slopes of the canyon are much oversteepened, the upper limit of glaciation being estimated at between 1500 and 2000 feet north of Allen Glacier and 1450 feet at Childs Glacier. Below the shoulder of oversteepening the slopes are rounded and there is little post-glacial stream dissection.

Above the shoulder of oversteepening (a–a in Pl. CLXXXI, A,) there are many stream courses, separated by rather finely dissected spurs. The peaks include many pointed horns and broad towers, especially in the horizontal strata just north of Allen Glacier and Wernicke River.

The great predominance of striation and glacial rounding and polishing upon the lower slopes of the canyon emphasizes the few surfaces of bare rock which are hackly and innocent of glácial grooving. Among the places where these are found are, (1) the pre-

[1] Tarr, R. S. and Martin, Lawrence, Glacial Deposits of the Continental Type in A'aska, Journ. Geol., Vol. XXI, 1913, pp. 289–300.

[2] Allen, H. T., Copper River, Alaska, Glacial Action, Science, Vol. VIII, 1886, pp. 145–146.

cipitous 1800 foot cliff east of the terminus of Allen Glacier, where the lower 1000 feet are nearly vertical, (2) the western canyon wall at Abercrombie Rapids, and (3) the western side of the isolated 600 foot rock hill at the mouth of La Gorce Valley. Part of this rock, facing Allen Glacier, has vertical and, at one place, actually overhanging cliffs. In each case the absence of striation, which is unquestionable at the localities inspected, and seems to be so for the inaccessible higher portions of the same cliffs examined with field glasses, is explained by glacial plucking. The absence of striation on two of these cliffs, however, opposite marginal channels of the Copper River, suggests the complication of stream erosion. Below these non-striated cliffs there is usually an absolute absence of talus, showing that the rough surfaces were produced by an erosive agent which worked in association with a transporting agency. This eliminates post-glacial weathering or avalanching.

Talus slopes mantle certain of the canyon walls in the section between Tasnuna River and Childs Glacier. These are especially well developed at Abercrombie Rapids, where the railway engineers have cut off the noses of several steep talus cones which are 1000 or 2000 feet high. This has modified them in a way to invite future rock avalanches.[1] These talus cones, some of which extend above the upper limit of glaciation, and which are now entirely covered with vegetation are, therefore, no longer growing, from which it may be inferred that post-glacial, high-altitude weathering is far less effectual than the super-glacial weathering of the time of maximum glaciation.

Some of the tributaries of Copper River are in hanging valleys (Pl. CLXXIX). The smaller tributaries have discordances of several hundred feet, the stream at Whiting Falls north of Heney Glacier, for example, hanging 300 to 400 feet, another (Map 9) just southwest of Allen Glacier hanging 550 feet, and a third just north of Abercrombie Rapids hanging about 400 feet. This last hanging valley is at Camp 55 and a milky stream from a small glacier descends its lip, but the gorge of the stream just north of it, which has no glacier, is not a hanging valley. Shields Glacier is in a hanging valley, but it is not certain whether the large tributary streams, like those of La Gorce Glacier and Wernicke River are in hanging valleys or not. If so, their discordance is hidden by the gravel filling of the Copper River canyon. Bremner River may even be discordant in relation to the Tasnuna and main Copper Rivers, for there are low, roches moutonnée hills in its mouth.

The floor of the middle section of the Copper River canyon is absolutely devoid of rock ledges, excepting in rare hills like the large one at the mouth of La Gorce Valley and small rock islands close to the valley walls, as in a case just south of Wernicke River. The rock bottom is thought to be some distance below the present river level, perhaps partly accounting for the greater width here than in the section north of Tasnuna River. The valley floor is mantled with glacial outwash, which covers about fifty square miles between Allen Glacier and Tasnuna River. The present grade of Copper River for the whole middle section of the canyon averages less than three and one-half feet to the mile, or about half that in the section between Tasnuna River and Chitina. As this three and one-half foot average includes the Childs, Abercrombie, and Baird Rapids, it is apparent that the grade elsewhere is not over 2 or 3 feet to the mile. The Allen (Pl. CLXXXI, B) and Miles Glaciers, acting as a dam and checking

[1] Some of these predicted avalanches took place during the summer of 1912, blocking the railway for several weeks.

the current of the Copper River, are responsible for this great accumulation of coarse, outwash gravels. The river is divided into innumerable, branching, shifting channels of the type which develop in heavily-loaded glacial streams whose current is checked (Pl. CLXXIX).

The Former Copper River Fiord. The third section of the course of the Copper River across the Chugach Mountains is between Childs Glacier and the railway bridge at Flag Point. This is the mouth of the Copper River canyon, which flares so much in a distance of twelve miles that the width of four miles at Childs Glacier increases to thirteen miles at Flag Point. The mountain walls attain heights of 4000 to 5000 feet within 1 to 3 miles of the valley bottom.

The lower portions of the valley wall slope very steeply as a result of glacial erosion. In a cliff just south of Miles Glacier on the eastern valley wall there is an ascent of 1700 feet in a distance of 2000 feet, and just north of Flag Point on the western valley wall there is an ascent of 4000 feet in less than a mile.

The tributary valleys enter at discordant grades, as near McPherson Glacier where extensive glacial erosion is shown by the hanging valley over the end of which this ice tongue cascades. The hanging cirque between McPherson and Miles Glaciers has been eroded by a local glacier now melted away. The valley in which Fickett Glacier lies seems to have been much deepened and steepened in the lower slopes by ice erosion. The Childs, Goodwin, and Johnson Glaciers probably occupy hanging valleys, but is not certain whether the valley occupied by Martin River Glacier is discordant or not.

It is thought that this lower, flaring portion of the Copper River canyon is a former fiord and that the waters of the Pacific Ocean formerly extended at least as far north as the Childs and Miles Glaciers.

This conclusion is based upon the fact that borings at the mouth of the canyon near Flag Point, close to the western valley wall, and at Hotcake Channel, about in the middle of the valley, show that there is no rock for a distance of at least 25 to 35 feet below sea level. At the head of the flaring section of the valley the soundings in the Miles Glacier lake show that the rock bottom is over 184 feet below sea level.

From these facts it may be concluded that when the trunk glacier of the Copper River canyon retreated it had deepened this flaring embayment so that the lower fifteen miles or so were below sea level. That the ice did not halt long enough to fill this embayment at once is suggested by the absence of recessional moraines. That the ocean then extended up to or above present sea level is shown by the presence of wave-cut forms, like the stacks at and west of Alaganik. If the water extended inland to the stacks at Alaganik it must have also filled the Copper River fiord, at least as far north as Miles and Childs Glaciers.

The flaring form of the valley walls indicate that this former fiord had outlines similar to Orca Inlet, Sheep Bay, Port Gravina and several other fiords to the west, which would be readily transformed to filled fiords exactly like the mouth of Copper River canyon if great glacial streams flowed into their upper portions.

The Copper River Delta. The Copper River has not only completely filled this former fiord, but it has built a delta (Pl. XCIII) which extends about 12 miles out into the sea south of Flag Point and shallows the Pacific Ocean for twenty miles more. This delta coalesces with the alluvial fans of Sheridan, Scott, and adjacent glaciers, on the

west and the outwash plain of Martin River Glacier on the east. The Copper River delta merges so imperceptibly with the outwash gravels which fill the former Copper River fiord that it is best thought of as heading at Childs and Miles Glaciers, especially because just below Goodwin Glacier the Copper River branches into a series of distributaries.

Details Regarding the Glacial Streams on the Delta. At present the Copper River distributaries occupy the western half of the delta and the Martin River distributaries the eastern half. The largest of the Copper River channels follows the extreme western side of the valley and at the lower bridges on the Copper River and Northwestern Railway there are smaller branches east of Long Island, the chief one called Hotcake Channel; and two main streams west of Long Island, separated by Round Island. We shall call the westernmost and largest of these the Flag Point Channel and the smaller or eastern distributary, between Round and Long Islands, the Long Island Channel. Round and Long Islands each rise 30 to 60 feet above sea level, or 12 to 48 feet above all stages of water, and there are numerous sand bars at their borders at the lower stages of the river.

The combined width of these three chief distributaries of Copper River is 4445 feet, in contrast with a width of 1500 feet in the single channel of the undivided river at Miles Glacier bridge, where the depth is not much greater but the velocity is several times that at Flag Point Channel. The relationships of width and depth and velocity are shown below, the total volume of the river having increased slightly below the Miles Glacier bridge by receiving water from the Childs, Goodwin, and smaller glaciers. In this table high water widths and depths are used, but the velocities are determined at various stages of river and are not so directly comparable.

TABLE SHOWING WIDTHS, DEPTHS, AND VELOCITIES OF COPPER RIVER ON ITS DELTA

Name of Channel	*Width of River in Feet*	*Depth of River in Feet*	*Velocity in Miles per Hour*
Miles Glacier Bridge	1500	33 to 40	6 to 12
Flag Point Channel	3200	6 to 29	2.8 to 3.6
Round Island Channel	525	39	3.2 to 3.4
Hotcake Channel	720	18	———

It may be of interest to compare and contrast the conditions at the head of the Copper River delta, where the river is in a single channel, with those at the three distributaries to the south. For this purpose data must be taken from (a) the channel at Miles Glacier Bridge and (b) at a point just south of Childs Glacier.

At Miles Glacier bridge, where there is a 1550 foot steel bridge, the river level varies from 116 to 140 feet above mean low tide at sea level, giving a vertical annual range of 24 feet. The channel is 8 to 16 feet deep at low water, when it is divided by a great gravel bar. At low water the velocity of the current is 3 to 3.8 miles an hour, but even

with the water at an elevation of 132.3 feet, or nearly 8 feet below the high water stage the velocity is 12.8 miles an hour.

The speed of the current increases rapidly at this bridge, where the water is coming out of the Miles Lake and starting down the rapids in front of Childs Glacier. At low water, for example, the velocity of 1.6 miles an hour about 1600 feet above the bridge increases to 1.7 miles in the first 500 feet, to 2.3 miles in the next 500 feet and to 3 miles an hour in the next 700 feet. At a medium stage of water the velocity increases from 3.7 miles an hour above the bridge to 6.1 miles an hour 800 feet down-stream below the bridge. A long series of velocity determinations were made by the railway engineers in the summer of 1908, by watching with two transits the rate of flow of icebergs in different parts of the river. These show that at a middle stage of the river the average velocity at this point is 6.3 miles an hour. By measuring the cross-section of the channel and the velocity of the current at various stages of water it is possible to compute the number of cubic feet of water which pass this bridge in a second. At high water this exceeds 300,000 cubic feet per second, equalling the summer volume of Mississippi River at New Orleans. At low water it falls below 50,000 cubic feet per second.

Detailed observations show that the river is highest in summer, when melting of the glacier ice is the greatest, and lowest in winter, when river ice generally covers the river from November to the last of May, or first of June.

At the site just south of Childs Glacier, where the building of a railway bridge was once considered, the channel is wider and shallower than at Miles Glacier Bridge. It was markedly shallowed by deposition between 1906 and 1908.

The river descends between 40 and 50 feet in the 3 miles of rapids from Miles Lake to the southern edge of Childs Glacier, a grade of about 15 feet to the mile. The grade is much flatter from this point to the bridge at Flag Point, averaging less than 4 feet to the mile.

The Flag Point crossing consists of a 700 foot and a 600 foot steel bridge west of Round Island, and two intermediate stretches of heavy trestles, crossing the sand bars which are exposed when the river is low. The river level at the Flag Point bridges varies from 10.25 to 18 feet above mean low tide at sea level, so that there is a vertical annual range of eight feet in the river level in the portion of these channels near the Flag Point bridges. Here detailed soundings and determinations of rate of stream flow were made in 1910 by A. O. Johnson, one of the railway engineers.

The velocity of the current in the Flag Point Channel was 3.4 to 3.6 miles an hour in August, 1910. Here the low water channel is narrowed to 600 feet, with an increase in velocity due to constriction, for the water only moves 2.3 miles an hour north of the bridge where the channel is twice as wide. The depth of the river at low water is eight to twenty feet, and portions of the channel were deepened from seven to eighteen feet, while other portions were shallowed slightly by deposition between August, 1908, and July, 1910, showing how rapidly the channels of a great glacial stream, heavily laden with sediment, may vary. This may be due in slight amount to the interference with normal conditions by the piers of the bridge, but doubtless it mainly represents the variations in the bottom profile of a glacial stream, for the sections between bridges were also markedly deepened between 1908 and 1910.

The Round Island Channel is crossed by a 560 foot steel bridge. Here the velocity varies from 2.8 to 3.4 miles an hour, the depth from 2 to 30.9 feet below low water,

and the channel was deepened 12 feet in some portions and filled 12 feet in others between August, 1908, and August, 1910, in this case clearly as a result of interference with the current by the piers of the bridge.

Six miles farther to the northeast where the railway crosses the eastern distributary of Copper River from Long Island to the eastern bank, there is a main stream, Hotcake Channel, and several minor ones. Hotcake Channel is crossed by a 200 and a 400 foot steel bridge and an intermediate trestle. This channel, 15 miles from the ocean, is from 1 to 8 feet deep at low water, the low water stage being 31 feet above mean low tide at sea level and the range of river level being about 7 feet.

Throughout this portion of the Copper River delta the detailed soundings afford an especially good opportunity of watching the future variations of a glacial stream, the shifting of sand bars as revealed at low water, and the cutting and filling in various parts of the channels, as they may be revealed by soundings near the railway bridges.

Coarse Deposits of the Glacial Streams on the Delta. The surface material in the outwash deposits of Copper River delta has already been spoken of as gravel, sand, and silt or clay. The same materials are found for a considerable depth below the surface. At the Miles Glacier bridge, for example, the river is in a 50 foot gorge and borings extend to a depth of 40 to 50 feet below the bed of the Copper River, revealing such a section as the following, in which the exposed cliff on the south bank of the Copper River supplies the upper 56 feet and the sinking of the caisson for Pier 2 of the railway bridge the remainder.

SECTION OF COARSE GLACIAL OUTWASH DEPOSITS AT MILES GLACIER BRIDGE

Material	*Thickness in Feet*
Bowlders, gravel and sand [1]	56
Large bowlders	3
Large bowlders, lying in cemented sand, gravel and silt	11
Large bowlders in loose sand and gravel	4
Occasional bowlders in loose sand and gravel	3
Bowlders in loose sand and gravel	2
Coarse, heavy sand and gravel	3
Compact sand and gravel, growing more compact at bottom	6
Total thickness revealed	88

Borings below the bottom of the river at the railway bridge across the Flag Point Channel reveal the following sections, all thicknesses being given in feet. All these deposits except the first in the left-hand column (sand and gravel, 18 feet) are below the high water surface of the river.

[1] From the surface of the plain of outwash gravels at Miles Glacier railway station to the surface of the gravel bar in the middle of Copper River.

Sections of Fine Glacial Outwash Deposits at Western Bridge across Flag Point Channel

Material	*Thickness*	*Material*	*Thickness*	*Material*	*Thickness*
Sand and gravel	18				
Sand	4				
Sand and gravel	4				
Unrecorded	6				
Fine sand	6½	Glacial mud	4	Glacial mud	7
(Sea level)		(Sea level)		(Sea level)	
Fine gravel	7	Fine sand	15	Fine sand and gravel	18½
Coarse gravel	17	Sand and gravel	18	Coarse sand and gravel	16
Silt and vegetation	18	Gravel	11	*Silt and vegetation*	8
Gravel	2			Sand and gravel	2
Sand	12				

The first point concerning these sections is that they show moderate-sized material, compared with the bowldery outwash, shown in the section (p. 461), 20 miles to the north at Miles Glacier bridge.

The three sections given above are 250 to 350 feet apart and the lack of correspondence of the sand or gravel layers shows the normal horizontal variation common to river deposits.

Eight other borings between this point and Long Island show similar relationships and variations, which indicate that, in the past, there have been the usual differences in velocity and carrying power of this glacial stream, which alternates in depositing sand, gravel and silt. These borings go 50 to 60 feet below sea level, but the total thickness of the glacial deposits of the Copper River delta is quite unknown. Six miles farther northeast at Hotcake Channel the borings show similar gravel and sand, extending below sea level.

Vegetation Indicating Sinking of the Land. The presence of vegetation in the silt at a depth of 20 to 40 feet below sea level in two of these borings shown in the table and in two others to the eastward shows that the Copper River delta has been subjected to a submergence in recent times, as is the case in parts of Prince William Sound to the westward, as well as 30 miles east near Controller Bay.

In the latter locality E. C. Hawkins states that in putting down borings, to see if it was possible to support a railway trestle on piles in the shallow Bering Lake, a layer of peat was encountered 30 feet below the surface. As the tide now rises into this lake the depth of this peat layer below sea level is about the same as on the Copper River delta.

Unfortunately the nature of the buried vegetation in the borings at Flag Point is not specified further, but, whether the vegetation was peat from a salt marsh or fresh-water marsh, or stranded tree trunks, it is clear proof of sinking of the land. It is inconceivable that water-lodged trees, for example, should happen to be buried in one layer that is now 30 or 40 feet below sea level in all four of the sections and that elsewhere in the deposits there should be none.

This proof that the last movement of this coast line was downward is of much interest because there has also been post-glacial uplift of this same coast, shown on Wingham Island forty miles to the southeast, where there are uplifted silts containing striated, glacial pebbles and marine fossils. [1]

Extent of Outwash Deposits. The deposits of outwash laid down by glacial streams on the Copper River delta itself, on the valley train of Martin River Glacier, and on the outwash plain of Sheridan and Scott Glaciers form an area of approximately 500 square miles. They vary in character from bowldery conglomerates and coarse gravels at the head of the delta near Childs and Miles Glaciers to fine sand, silt, mud, and clay in the outer Copper River delta. They vary also in shape, the upper portion of the Copper and Martin River deposits forming a smooth outwash plain, the lower deposits of the same rivers forming a smooth delta, while the outwash gravels of the Sheridan, Scott and adjacent glaciers form an outwash apron of coalescing alluvial fans.

Valley Train of Martin River. The outwash deposits of the Martin River Glacier west of the terminus of that glacier, have a length of 10 miles and a width of from 3½ to 5 miles up to the point where they join the outwash deposits of Copper River valley. They are densely forested, where stream deposition is no longer going on, but barren along the present streams. The course of the Martin River, which, in 1905, emerged from the glacier near the northern side, joined by braided streams from a small cascading glacier, curves southward across the whole valley, because of the presence of a recessional moraine three miles west of the present terminus of the glacier, which interrupts the even slope of the outwash gravels. The present position of Martin River, close to the southern side of its valley, has resulted in the building up of the valley train higher than the mouths of five or more tributary valleys, in each of which the water is ponded back, forming good-sized lakes which have been described by G. C. Martin.[2]

Alluvial Slope of the Western Edge of the Delta. The outwash deposits of Sheridan and Scott Glaciers form a confluent outwash plain with so even a slope that the Copper River and Northwestern Railway has been built across it with a perfectly straight stretch of track for over eleven miles. The surface rises and falls, however, in a series of broad, low, alluvial fans, but the track is built over the undulating surface without much grading. Along the line of the railway the highest point is 49 feet above sea level, and the lowest point 15 feet, so that the maximum relief is 34 feet. These fans are highest near the present streams, or the places where large streams formerly flowed for a long time, the differences in level between the axis of an individual alluvial fan and the depression on one side being 34 feet in 2 miles, 11 feet in 1 mile, etc., according to the line of the section with reference to the head of the fan.

There are five main alluvial fans on the outwash gravel plain of the Sheridan and

[1] Martin, G. C., Bull. 335, U. S. Geol. Survey, 1908, pp. 46, 64.

[2] Bull. 335, U. S. Geol. Survey, 1908, p. 56.

Scott Glaciers whose relationships, where crossed by the railway, are shown in the following table. This table shows clearly that the glacial streams abandon the higher points of their alluvial fans when they are built to a certain height, for at the time this railway leveling was done only one of the five fans had a stream upon its crest and that was a very small one.

TABLE SHOWING A CROSS-SECTION OF ALLUVIAL FANS WEST OF COPPER RIVER DELTA

Position	*Elevation in Feet*	*Position on Fan*	*Distance in Miles*	*Ascent or Descent in Feet*
Eastern alluvial fan				
East Sheridan stream	15		—	—
Small stream at Mile Post 17	49	Crest	2	+34
Near Mile Post 15	32.9		2	−16
Second alluvial fan				
Near Mile Post 14	43.9	Crest	1	+11
Main Sheridan Stream	32.9		$\frac{1}{2}$	−11
Near Mile Post 13	31.5		$\frac{1}{2}$	$-1\frac{1}{2}$
Third alluvial fan				
In Mile 12	42	Crest	$\frac{1}{4}$	$+10\frac{1}{2}$
Near Mile Post 11	28		$1\frac{3}{4}$	−14
Fourth alluvial fan				
Near Mile Post 10	35	Crest	$\frac{1}{2}$	+7
Main Scott Stream	16		$1\frac{1}{2}$	−19
Fifth alluvial fan				
Middle of Mile 7	20	Crest	$\frac{7}{8}$	+4
At Mile Post 7	15		$\frac{1}{2}$	−5
Middle of Mile 6	12.3		$\frac{1}{2}$	$-2\frac{1}{2}$
Outlet of Eyak Lake	19.7		$\frac{1}{2}$	$+7\frac{1}{2}$

This outwash gravel plain grades to fresh marsh, then to salt marsh, but in the area between the Copper River and the Sheridan Glacier streams the salt marsh extends up to the mountain base.

Relation of Outwash Deposits to Maintenance of the Railway. Two points stand out in connection with the building of the Copper River and Northwestern Railway on the Copper River delta. One is the ease of original construction upon these smooth alluvial fans, delta plain, and valley train, aside from the necessary filling in swamps and on quicksands. The other is the threat of expense of maintenance of the railway.

The danger that the Sheridan Glacier might advance two and one-half miles and destroy the railway now seems practically negligible, but the danger that it may advance slightly and that the streams from it may shift and necessitate new trestles or may bury the track in places, as at Spencer Glacier on the Alaska Northern Railway,[1] is an immi-

[1] Tarr, R. S. and Martin, Lawrence, An Experiment in Controlling a Glacial Stream, Annals Assoc Amer. Geographers, Vol. II, 1912, pp. 25–40.

ICE STRUCTURE IN ALLEN GLACIER
Folds of a type characteristic of incompetent materials like those of the zone of rock flowage.

PLATE CLXXVIII

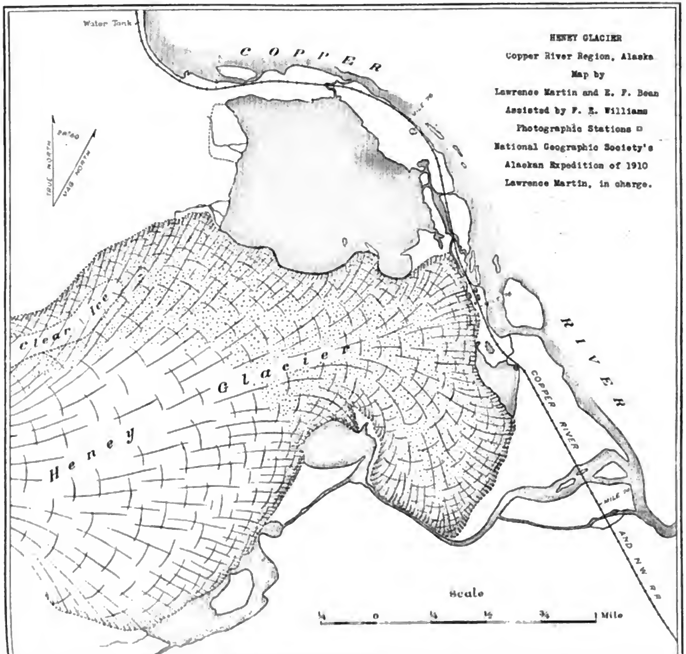


PLATE CLXXIX

MAP OF THE BROAD MIDDLE PORTION OF THE COPPER RIVER CANYON
Between Tasnuna River and Allen Glacier. (After U. S. Geological Survey.) Hanging valleys shown by letter h.

Plate CLXXX

A. Furrowed Surface from which Heney Glacier had Recently Retreated in 1910
Southern terminal moraine in the background.

B. Details of Furrows and Ridges Shown in Upper Illustration

A. VIEW IN THE BROADER MIDDLE PORTION OF THE COPPER RIVER CANYON
Upper limit of glaciation near a, a, below which the slopes are smoothed by glacial erosion. Photograph by E. A. Hegg.

B. THE COPPER RIVER IN ITS MARGINAL CHANNEL
Precipitous eastern wall of canyon on left and Allen Glacier on right.

PLATE CLXXXII

GLACIAL LAKES
On lower, moraine-covered slope of Hinchinbrook Island. Photographed by White.

Plate CLXXXIII

Hinchinbrook Island
Level upland of driftless area.

Plate CLXXXIV

Sketch Showing Former Extent of Glaciation in the Prince William Sound and Lower Copper River Regions

nent one. Some measure of annual expense for new trestles and for raising the grade is to be looked for, even if the adjacent glaciers remain stationary or retreat slowly.

The second threat of extra expense of maintenance comes from the Copper River between the Flag Point and Hotcake Channels. There have been changes in the stream bed in even the short time since the bridges were built. While the steel spans were being erected the temporary trestles beside them constantly had to have new piles driven, because the current had undermined them. The steel bridges, built on concrete piers, seem safe, though not as safe as if it had been possible to erect the piers upon solid rock, rather than upon piling. The chief danger alluded to, however, is not that of undercutting by erosion, but of stream deposition. Unless artificially confined, the time will come when the several, heavily-loaded, Copper River distributaries will shift out of their present channels as they are silted up, leaving the expensive steel bridges high and dry, and necessitating new bridges wherever the new channels establish themselves. The railway engineers have to deal with what is to them a new type of stream and even the experience in attempting to control the lower Mississippi and in correcting small glacial streams in the Alps will fail to furnish adequate data as to what a source of trouble and expense a glacial stream may be to a railway and how it may best be coped with.

Fine Glacial Deposits of the Outer Edge of the Delta. We lack detailed information regarding the nature of the Copper River distributaries between the railway bridges and the sea. It is thought, however, that the grade continues to flatten, resulting in the deposition of the coarser silt carried in suspension by the river and causing the even shallower channels and tidal sloughs south of Flag Point, where the river is navigable only by small boats.

The bottom of Copper River is from 10 to 20 feet below sea level at the Flag Point and Round Island Channels, though slightly beyond the range of the tides and over twelve miles from the ocean.

Upon the edge of the Copper River delta the Coast Survey charts[1] show a horizontal range of the tide of several miles. Offshore from the Scott and Sheridan Glaciers, for example, continuous tide flats are exposed for distances of from 5 to 7½ miles at mean lower low water. The vertical range of the tide is only ten feet, showing that the slope is very slight indeed. Crossing these tide flats are small channels or sloughs, navigable only by launches. More than two miles from shore at the eastern edge of the delta, near the mouth of Martin River, there is a depth of only 1½ to 3 feet at mean lower low water. Opposite the middle of Copper River delta there are depths of only 54 to 102 feet nearly nine miles offshore, beyond which the water deepens to 414 feet in less than two miles. The bottom material is sticky mud, and the Pacific Ocean is discolored for over twenty miles offshore by the glacial silt in suspension, part of which goes westward and has helped fill Orca Inlet. The nature of the fine glacial sediment deposited on the outer part of the delta and exposed on the mud flats at low tide is well indicated by Seton Karr,[2] who tells of his experience in crossing the border of the delta in 1886 in a canoe with several natives. They were traveling inside a line of sandy islands, which furnished protection from the waves of the Pacific, and when the tide was too low to float the canoe any longer they were able to continue by sliding it over the slippery surface of the tide flats, continuing to sit in the canoe and push it

[1] U. S. Coast and Geod. Survey, Charts 8502, 8520, and 8513.

[2] Seton Karr, H. W., Shores and Alps of Alaska, London, 1887, pp. 167–168, 171.

along with the paddles while a little water remained and, after that, dragging it easily over the bare, slippery surface of the fine glacial sediment.

Wind-Blown Glacial Deposits on the Delta. In a number of places sand dunes rise above the Copper River delta to a considerable height. The material is sand formerly carried out by glacial streams. There are large sand dunes 30 to 40 feet high on Long Island, just east of the Round Island Channel, which necessitate sand sheds along the railway. There is another large area of sand dunes in the 150 foot hill visible to the south of Long Island. There are also sand dunes east of Hotcake Channel, as well as on Egg Islands and other sand bars to the eastward. The latter have been built up by the surf in the shallow water off the delta front. From Softuk Bar, near Katalla, to the mouth of Orca Inlet, these sandbars form an incomplete barrier 5 to 30 feet high and 4 to 7 miles off the front of the Copper River delta. While most of this sand comes from the Copper River and adjacent streams, some is doubtless supplied by the glacial streams from the Bering Glacier in Controller Bay and is moved along shore by the westward-setting currents.

Although a normal delta in many respects, and with normal deposition by water, modified by the rise and fall of the tide, by alongshore currents, and by wind work, the Copper River delta is conspicuous as the terminus of a great area of outwash deposits, brought by the great volume of sediments from the present-day and former glaciers of this region.

The Driftless Area Near Alaganik. An unforeseen feature in connection with the former expansion of the glaciers near the Copper River delta and the erosion accomplished by them, is the absence of any indication of glaciation along parts of the mountain front between Sheridan Glacier and the Copper River. At several points along the railway, as near Alaganik, there are rock pinnacles standing out from the mountain front and resembling stacks along a sea shore. Such forms as these are easily overturned by a glacier, and it seems quite probable that they are stacks produced by the waves of the Pacific before the Copper River delta was built. They cannot be considered shore forms produced after the retreat of the glacier, for there is no explanation of the failure of the waves to remove the evidences of glaciation at other points on the coast and higher above sea level than this.

The conditions in the glaciated areas adjacent to the driftless region are unmistakable. That the ice expanded for some distance outside the mountain front in the region west of this driftless area is shown by the fact that the lower rock ledges in the vicinity of Sheridan Glacier bear striae, and there are many roches moutonnée rocks, like the one at the front of Sheridan Glacier: there are extensive cirques in the mountains west of Scott Glacier, where there is a hanging valley half way between Scott Glacier and the railway; the valley containing Eyak Lake, between Scott Glacier and Cordova, was much deepened and had its lower slopes steepened by the expanded southern tongue of Shephard Glacier, which filled the whole valley of Eyak Lake; along the railway, striae on the rocky spur projecting into Eyak Lake from the south show that the ice tongue split on this point, sending one glacial distributary southwestward past Cordova, and the other southeastward to join the expanded Scott Glacier. All of the glacial drift between Eyak Lake and the driftless area at Alaganik is outwash, except for the eastern terminal moraine of Sheridan Glacier which contains till and abundant granite bowlders.

East of the driftless area the conditions are likewise unmistakable. The oversteepened western wall of the former Copper River fiord between Fickett Glacier and Flag Point shows that the trunk glacier of the Copper River valley extended south of Flag Point. Saddlebag Glacier valley also shows unmistakable evidence of former expansion as does the valley of Salmon Creek and McKinley Lake.

Between these regions is one where absence of striae on the rock ledges and of erratics in the soil seems to point to absence of invasion by glacial ice. The incoherent material forming the flat floor of the region is all fine silt. The new railway cuts afford ample opportunity to find foreign bowlders, but none were found. The western boundary of the driftless area is well-defined. On the railway it is almost eight miles west of the Copper River, where a well marked terminal moraine of Sheridan Glacier crosses the track, 625 feet east of Mile Post 19. The eastern boundary is less certain. No signs of glaciation are visible along the railway in the eight miles eastward to Flag Point, and it is uncertain how far outside the mountain front the Saddlebag and McKinley Lake Glaciers expanded.

The cause of this driftless area was that the expanded Scott and Sheridan Glaciers, which apparently united to form a piedmont ice tongue, probably extended southward far enough to coalesce with the trunk glacier from the Copper River canyon. In the angle between these glaciers and the mountain front there is a small driftless area, not overridden by the maximum glacial advance.

Vegetation and the Former Advance of Glaciers in the Copper River Canyon. Besides the interesting relationships of trees and shrubs growing upon the dirt-covered ice of Miles, Grinnell, Allen, Heney, and several other glaciers, as already described, the vegetation in the Copper River region tells a specific story of its recent glacial history. The Copper River delta and the southern part of the canyon support a dense coniferous forest, except where treelessness is caused (a) by ocean water on the tide flats and salt marshes, or (b) by aggrading glacial streams on the outwash gravel plains. The spruces are abundant from the ocean up to McPherson Glacier, but not north of there, though there is a thick forest of cottonwoods and alders, and one baby spruce was seen near Miles Glacier bridge. From this point northward no conifers were observed in the Copper River canyon south of the Tasnuna and Bremner Rivers. North of Tasnuna River, however, there are spruces along the canyon bottom and its lower slopes, but in the Copper River basin north of Wood Canyon and near Chitina there is thick, mature, spruce forest.

The distribution of vegetation, therefore, shows a practical absence of spruces and hemlocks for nearly 40 miles along the Copper River, while at either end there is dense, mature, coniferous forest. The 40 mile strip which has no conifers is not barren, but is thickly covered with cottonwood and alder, wherever the slopes, the glaciers, and the glacial streams permit. Throughout the region the vegetation is terminated upward by a normal timberline.

To account for this distribution we postulate a recent period of expansion of the glaciers between Tasnuna River and McPherson Glacier. The retreat from this expanded stage was long enough ago for occupation of the 40 mile strip by alders and cottonwoods and for a commencement of occupation by the conifers, as in the case of the young spruce near Miles Glacier bridge.

Mr. W. G. Weigle of the U. S. Forest Service has stated[1] that "the absence of timber throughout upper Glacier Bay near the Muir Glacier and in many other places in southeast Alaska is so evidently the result of recent glaciation that I have been prone to attribute the absence of timber on the Copper River to the same cause, yet there may be other causes. One possible cause of this absence of timber worthy of consideration is that the spruce found near the mouth of the Copper River is tideland spruce (*Picea sitchensis*) and is only found near tidewater where there is a heavy rainfall. The spruce found on the Bremner, Tasnuna, Chitina and other portions of the upper Copper River basin is known as white spruce (*Picea canadensis*) which is one of the spruces found throughout the interior and other portions of Alaska but is not found in the excessively wet belt along the coast from Seward south. Another condition which may largely be responsible for the absence of timber in this region is that the Childs, Miles and Allen Glaciers even yet form a great barrier between the timbered regions of the upper and lower Copper River and the indications are that in quite recent times this ice field was much larger than it is at present, hence formed a greater barrier."

It seems probable that this distribution is due to an actual readvance of the ice tongues in the forty mile strip, rather than a lingering of glaciers there. This is shown by the thick, mature forest north of Tasnuna River and in the Copper River basin near Chitina. Here all forest was certainly removed at the stage of maximum glaciation, for the ice was thick enough to extend above the timberline. We think it unlikely that the conifers readvanced to the Copper River basin by way of the deglaciated Copper River canyon because of the difference of species described by Weigle. It seems more probable that the unglaciated territory to the westward near Cook Inlet supplied this northern region with conifers. The lingering of conifers at the mouth of the Copper River canyon during the stage of maximum glaciation was made possible by the leaving of the driftless area near Alaganik and of driftless areas to the west on Hinchinbrook Island, and probably also to the east in the Controller Bay coal field, where G. C. Martin[2] states that at the maximum stage of Bering Glacier the ice rose only 200 to 700 feet above the present ice surface, which we interpret as below the timberline of that period.

The forty mile strip from which the conifers are absent is near the largest glaciers, and the present vegetation of the Copper River canyon is in an intermediate stage between the Prince William Sound region, where mature forest extends up close to the present glaciers, and the Yakutat Bay region, where the readvancing forest near the ice fronts is even less mature than in the Copper River canyon. Apparently the Heney, Allen, Miles, Childs, Goodwin, and other glaciers of this part of the Chugach Mountains advanced and destroyed the vegetation in part of the Copper River canyon at some fairly recent period. We should not know of this if it was not for the absence of conifers in the forty mile strip between Tasnuna River and Childs Glacier.

Glaciation of the Lower Copper River Region. The present stage of glaciation of the lower part of the Copper River valley is one where the larger side valleys are filled with ice tongues, projecting into the main valley and expanding there as bulbs which obstruct the major drainage and supply Copper River with great amounts of water and sediment. A former stage, perhaps repeated, involved the occupation of this main valley by a

[1] Personal communication, November 21, 1912.

[2] Martin, G. C., Bull. 335, U. S. Geol. Survey, 1908, p. 50.

great trunk glacier (Pl. CLXXXIV) to which the expanded Miles, Childs, Allen, Heney, and other glaciers were tributaries. This is indicated by the hanging valleys, as well as by oversteepened valley walls, eroded spurs, and the broadened valley bottom.

We realize that many problems are left unsolved, and, of these, one of the most interesting is the behavior of the Copper River during earlier stages of glaciation. At present, with its vast volume of water and its steep grade, it is able to compete with the glaciers and maintain mastery, though here and there diverted and transformed to rapids in the struggle, and in parts of its course burdened with a load of floating ice to be borne away. In earlier stages the struggle for mastery between river and glaciers must have been different, and some records of this struggle remain, but their exact interpretation calls for a much more detailed study than we were able to give.

These glaciers on the lower Copper River, and the glacial phenomena, are of more than ordinary interest, because of the fact that they lie along one of the highways to interior Alaska which promises to be of increasing importance in the immediate future. They will be easily accessible to the traveller from now on, and it is largely for this reason that we have devoted so much time to the glaciers and glaciation of the Copper River region, so that others may continue the work which we have merely commenced.

Prince William Sound

The Piedmont Glacier of Prince William Sound. It is clear that at the stage of maximum glaciation Prince William Sound was occupied by a great piedmont glacier. It was fed by the expanded valley glaciers of the fiords between Port Valdez and Orca Inlet on the east, by the enlarged ice tongues of Columbia Bay, Unakwik Inlet, Port Wells, and other fiords on the north, and by the Passage Canal, Port Nellie Juan and Icy Bay Glaciers on the west.

Its general extent is shown in Pl. CLXXXIV, a generalized map upon which no attempt has been made to show the complete seaward extent of this piedmont glacier.

It covered all of the lower islands and rose high upon the slopes of the tributary fiords, but, on some islands, high peaks rose through it as nunataks. Naked and Peak Islands, which are 1200 feet high, are shown by their rounded summits to have been completely overridden, as were Green and Perry Islands and the more irregular Lone Island. The nunataks were on Montague, Knight, and Hinchinbrook Islands where there are many horns and needles among the summit peaks.

The surface of the piedmont glacier may be restored in part, as is shown in the following table, in which some of the determinations of approximate heights of glaciation were made by Gilbert, some by Grant and Higgins, and some by the National Geographic Society's 1910 expedition. All altitudes are given in feet.

Table Showing Heights of Glaciation in Prince William Sound

Harriman Fiord	4000	Latouche Island	2000
Columbia Bay	4000	Hinchinbrook Entrance	400
Port Valdez	3200	Orca Inlet	2300

The points in the first column are in northern Prince William Sound, those on the right hand are near the Pacific Ocean, and Hinchinbrook Entrance is about midway

between Latouche Island on the west and Orca Inlet on the east. The slope of the former ice surface between the northern and southern points of determination is (1) about 25 feet to the mile from Harriman Fiord to Latouche Island, the grade flattening in the southern portion and being only about 17 feet to the mile between Knight and Latouche Islands; (2) about 60 feet to the mile between Columbia Glacier and Hinchinbrook Entrance; and (3) about 50 feet to the mile between Orca Inlet and Hinchinbrook Entrance. These grades are not quite as steep as those of the present Malaspina Glacier and of the ancient ice surface of Yakutat Bay, which average 70 to 75 feet to the mile. It will be noted that the surface of the piedmont glacier not only sloped southward from the northern snowfields toward the Pacific Ocean but also inward from the eastern and western snowfields to a low axis in the middle of the sound. This low axis was occupied by an ice stream which probably moved with some little rapidity, because it led to an outlet at Hinchinbrook Entrance.

Basins and Submarine Channels. The bottom of Prince William Sound, 200 to 400 fathoms deep, has been sculptured into basins and submarine channels by the movement of ice currents of the former piedmont glacier.[1] These channels show a direct relationship to the tributary fiords. Our knowledge of the submarine topography is based upon soundings by the U. S. Coast and Geodetic Survey, supplemented by our own, the data being complete except for the northwestern corner of the sound between Knight Island and the mouth of Passage Canal—Port Wells.

The largest channel (Fig. 72) leads from Columbia Bay and Valdez Arm to Hinchinbrook Entrance. It is about forty miles long and 7 to 10 miles wide, with an average depth of 1200 to 1500 feet. As the adjacent portion of the sound has a depth of only 400 to 700 feet it is evident that this channel is about 800 feet deeper than the portion of the sound through which it passes. At its head the expanded ice streams of Valdez Arm and Columbia Bay formerly came in from the northeast, joining a western ice stream from between Glacier Island and the Naked Island group, where the Wells Bay and Unakwik Inlet Glaciers evidently supplied part of their ice.

Ice streams from Port Fidalgo and from Port Gravina—Orca Bay came into the channel from the east and it is these that furnish proof as to the origin of this submarine channel. Port Fidalgo has a hanging valley relationship to this channel as does the united Port Gravina—Orca Bay. The different amounts of discordance, Port Fidalgo between 600 and 700 feet, Orca Bay 900 to 1000 feet, shows that these submerged hanging valleys are due to glacial erosion and that the great submarine channel itself is the result of ice sculpture. It also has the ice-eroded basin character, being 400 feet shallower near Hinchinbrook Island than it is a few miles farther north.

West of this channel and of northern Montague Island is a submarine ridge about ten miles wide and with water averaging less than 400 feet in depth. Above this ridge rise Naked and adjacent islands, Green Island, and a few small reefs.

A second submarine channel leads southward from between Naked and Knight Islands to Montague Strait, following close to the eastern shore of Knight Island. It is about thirty-five miles long and three miles wide and is 400 to 800 feet deep. Its relative narrowness and shallowness is due to the small amount of ice which moved down it, most of which probably came from Unakwik Inlet and Eaglek Bay and from local glaciers on Knight Island. Its glacial origin is shown by the fact that its northern end hangs

[1] Bean, E. F., Fiords of Prince William Sound, Alaska, Unpublished thesis, University of Wisconsin, 1911.

600 feet above another channel west of Knight Island, as well as by the submerged hanging valleys which it receives as tributaries from the eastern side of Knight Island. Of these, mention may be made of Snug Harbor, Marsha Bay, and Bay of Isles. Within the latter, are tributary coves, forming submerged hanging valleys of the second order.

A third channel beneath the surface of Prince William Sound lies northwest of Knight

FIG. 72. SUBMARINE TOPOGRAPHY OF PRINCE WILLIAM SOUND.

Island. The southern end of it leads into a fiord called Knight Island Passage. Near the mouths of Passage Canal and of Port Nellie Juan we lack soundings, but since the junction of Passage Canal and Port Wells is over 1400 feet deep and the northern known portion of this submarine channel is 2400 feet deep, it seems likely that in the intervening seventeen miles there is either a deep channel or several deep basins between the islands.

The ice in this northern portion of the channel was supplied by the expanded glaciers of College and Harriman Fiords, Passage Canal, Blackstone Bay, and Port Nellie Juan. It seems probable that at the maximum stage, Esther, Culross and adjacent smaller islands were completely overridden by the glacier.

The second portion of this channel is the deep basin north of Knight Island. It is at least twelve miles long and 2000 to 2400 feet deep, including not less than thirty square miles with a depth of 2400 feet. South of this basin the depth decreases to less than 1200 feet, suggesting that the depth of 2400 feet is due to glacial scooping caused by the obstruction to free outflow of the ice through the narrower channels to the southward.

Fiords and Islands of Western Prince William Sound. The fiord which connects directly with this submarine channel is northern Knight Island Passage. It is twenty-two miles long and 3 to 4 miles wide, with a depth of 800 to 2000 feet. When the glacier rose at least 400 feet above the crest of Chenega Island on the west and to a height of 2400 feet [1] on Knight Island to the east, the depth of moving ice in this channel was at least 3500 to 4500 feet. We have little hesitancy in ascribing tremendous erosive power to this ice stream and believe that it produced the whole fiord, which has steeply-sloping walls and receives tributary bays like Drier Bay, which hangs over 1000 feet and contains submerged hanging valleys of the second order like Cathead and Mallard Bays which hang 214 and 264 feet respectively.[2]

The eastern border of this fiord on Knight Island shows much influence, both of local glaciers and of the ice which formerly filled the fiord. At the time of maximum glaciation in Prince William Sound, a few of the higher peaks of Knight Island probably rose above the ice sheet as a group of nunataks, for Grant and Higgins state[1] that the island does not seem to have been glaciated higher than 2400 feet. The northern end of the island has rounded summits, but further south these give place to sharp peaks, with talus slopes and cirques containing snow, some of which may have been sharpened by the action of the local glaciers and by super-glacial weathering, perhaps after having been rounded by the erosion of the Knight Island Passage ice stream. Glacial erosion has been effective in rounding the predominating greenstones of Knight Island and producing a more knobby topography than on the slate and graywacke which make up Chenega Island, the mountains of Icy Bay, and other parts of the mainland to the west. There are U-shaped troughs and well-developed cirques and hanging valleys on Knight Island.

Within Knight Island Passage the shallowest portion of mid-channel, 800 to 1200 feet, is at the northern end of Chenega Island where a glacial distributary formerly moved southward through Dangerous Passage, west of Chenega Island. The deepest portion, over 2000 feet, is southeast of Chenega Island where a smaller island and the incoming of the expanded glacier of Icy Bay confined the ice tongue of Knight Island Passage to a narrow channel, forcing it to flow rapidly and erode deeply.

The southern portion of Knight Island Passage extends southeastward from Chenega Island to Montague Strait near Latouche Island. It is a fiord 4 to 6 miles wide and 15 miles long, exclusive of the mouth of Icy Bay. The depth ranges from 800 to 1200 feet.

[1] Grant, U. S. and Higgins, D. F., Bull. 443, U. S. Geol. Survey, 1910, p. 19.

[2] Charts 8550, 8515, and 8524, U. S. Coast and Geod. Survey.

The cross-section of this fiord is that of a steep-sided, flat-bottomed glacial trough, with tributary hanging valleys. A typical submerged hanging valley on the northeastern side lies beneath the waters of Mummy Bay on Knight Island. The bottom of Mummy Bay slopes from 126 feet at the head to 522 feet at the mouth, 3½ miles distant. The depth then increases to 1260 feet in Knight Island Passage, three-quarters of a mile outside. The slope above this hanging valley lip is 126 feet to the mile, while that outside is at the rate of 984 feet to the mile. Little and Hogan Bays near the southern tip of Knight Island also have submerged hanging valley relationships to the main fiord.

These relationships are of especial importance in connection with the origin of the southeastern portion of Knight Island Passage and outer Icy Bay which Grant and Higgins[1] have explained as due to graben faulting. We dissent from this view and ascribe the straight, steep-walled portion of Knight Island Passage east of Icy Bay and south of Knight Island entirely to glacial erosion. There is no published evidence of faulting, but the passage has every characteristic of an ice-sculptured fiord. The critical features, however, are the submerged hanging valleys. Such forms could be produced by faulting, it is true, but the different discordances of the lips of the submerged hanging valleys of different sizes cannot be explained by faulting, but are perfectly normal results of differential glacial erosion.

The depth of this southeastern portion of Knight Island Passage is less than might be expected in view of the incoming of the expanded glacier of Icy Bay, but is explained by the loss of ice through the distributary channels on the southwestern side of Knight Island Passage.

Latouche, Elrington, Hoodoo, and Flemming Islands, between Knight Island Passage, Montague Strait and the mainland, are separated from each other and from the mainland by narrow fiords called Latouche, Elrington, and Prince of Wales Passages, and Bainbridge Passage—Port Bainbridge, the four distributary fiords which diverted part of the ice of Knight Island Passage from Montague Strait. Latouche Island is a mountainous island but its coast is not so indented by minor bays and coves as Knight Island is. This may be due to the difference in the country rock, Knight Island being predominating greenstone with subordinate amounts of conglomerate, graywacke and slate, while Latouche Island is mostly slate and graywacke. It is separated from Elrington and Hoodoo Islands on the west, by a deep straight fiord. Latouche Island has its greatest elevations within ½ to ¾ of a mile of the southeastern side.

Grant and Higgins[2] have stated that Latouche Island was probably nearly or quite overridden by the ice sheet of Prince William Sound. The intensely glaciated, steep, southeastern side of the island forms a striking contrast with the cirque-carved, northwestern side previously described.

Elrington, Hoodoo, and Flemming Islands are each about twelve miles long. All of them, with the exception of Flemming Island, are narrower than Latouche Island, and more irregular, probably because the constituent rock, the same as in Knight Island, has allowed more dissection by stream and ice erosion. They also contrast in irregularity with the peninsula between Bainbridge Passage and Icy Bay where there is graywacke and slate, though of a different age from that in Latouche Island. Elring-

[1] Bull. 443, U. S. Geol. Survey, 1910, p. 16.

[2] Op. cit., p. 19.

ton and Hoodoo Islands arise to heights of from 1000 to 1967 feet, and Flemming Island is still higher.

None of these islands have glaciers. Grant and Higgins state[1] that the Prince William Sound Glacier extended nearly or quite to their summits, reaching about 2000 feet on the border of Port Bainbridge, and extending out past the end of Elrington Island, the southwesternmost island of Prince William Sound. This western distributary of the Prince William Sound Glacier, therefore, ended in the open Pacific Ocean.

We have not seen these islands except from a steamer in Elrington and Latouche Passages in 1904 and 1911, and from a launch in Knight Island Passage and eastern Latouche Passage in 1910; but the steep lower slopes give an impression of intense glaciation, being most abrupt on the southeastern side, as on Latouche Island.

These steep slopes are continued below sea level, as the Coast Survey charts show.[2] These fiords all follow the strike of the rocks and are all rather simple in outline, and flat-bottomed. The depth of Bainbridge and Prince of Wales Passages is unknown, except in the southwestern end of the latter where it reaches 504 to 606 feet. This contrasts with a maximum of 234 to 342 feet in the southwestern end of Elrington Passage, which is tributary to lower Prince of Wales Passage. The descent of 260 to 270 feet in a half mile of the fiord bottom suggests a submerged hanging valley relationship. This would be a natural result of greater glacial erosion in the broader Prince of Wales Passage, which leads straight from Knight Island Passage, and which doubtless had a more effective ice stream than the narrower, crooked Elrington Passage.

Within Elrington Passage there seems to have been differential glacial erosion, for this fiord, which has a right-angled turn near the southern end, reaches its greatest depth in a basin 594 feet deep, just at the elbow. This is an increase of depth amounting to 246 feet in a mile and a quarter, measured from the northeast, and 294 feet in five-eighths of a mile, measured from the northwest, and suggests that the ice tongue of Elrington Passage eroded this deeper rock basin because its free movement was retarded at this sharp turn.

Latouche Passage has two rather different parts. From Elrington Passage northeastward the depths vary from 480 to 750 feet, contrasting with the southwestern half, which is from 60 to 264 feet deep. This contrast suggests the possibility that before and after the main period of glaciation, when the ice probably streamed southwestward through Latouche Passage, the local glaciers from the great cirques of Latouche Island, most of which face the deep northeastern half of the fiord, extended out into the passage and supplied an ice stream which moved toward the northeast rather than the southwest, therefore eroding the broader northeastern part of the fiord most deeply.

There are submerged barriers on the bottom at both ends of Latouche Passage, the one facing the Pacific Ocean, near Danger Island, having a roughly-crescentic shape, with depths of from 3½ to 60 feet, in contrast with 264 feet just inside and 150 feet an equal distance outside. Two narrow channels with depths of 78 and 84 feet respectively cross this obstruction, which curves both ways from the low Danger Island. The shallow mouth of this fiord, contrasting with the deep, open termini of the wider Prince of Wales Passage and Montague Strait may be due either to lack of glacial erosion here or to the deposition of a terminal moraine below sea level.

[1] Bull. 443, U. S. Geol. Survey, 1910, p. 19.

[2] U. S. Coast and Geod. Survey, Charts 8522 and 8515.

The bar at the northern end of Latouche Passage has depths of 168 to 330 feet in mid-fiord, in contrast with 636 to 642 feet a half mile distant on each side. It is narrow and straight, completely closing the mouth of the fiord excepting for a 480 foot channel north of the middle. It may be explained equally well as either (a) a moraine bar or (b) an uneroded rock ledge with a rock basin southwest of it, though its position is not as favorable to the latter interpretation as in the case of the bar at the seaward end of the same fiord. If it is a submerged moraine it might have been built at the terminus of a southwest-retreating local glacier, fed by the cirques of Latouche Island, or as a lateral moraine of the northwest-retreating ice tongue of Knight Island Passage.

The Ice-Sculptured Outlets of Prince William Sound. The ice of the piedmont glacier of Prince William Sound had three outlets to the Pacific Ocean, (1) Montague Strait and the four adjacent fiords alluded to in the last section, (2) Hinchinbrook Entrance, and (3) the mouth of Orca Inlet.

The outlet by Montague Strait, was probably the largest one. It was not as important, however, as it would have been had there not been four subordinate outlets. This is shown by the fact that this outlet was only eroded to a maximum depth of 1100 feet and shallows to 400 feet near the Pacific Ocean. Its width is from 5 to 6 miles and the deepest channel is close to Latouche Island, perhaps because of the influence of local glaciers on Montague Island. The southeastern slope of Latouche Island is, therefore, smoothed and much oversteepened, sloping at the rate of 3500 to 4000 feet to the mile on the land. The slope continues below water at the rate of 1500 feet to the mile, beyond which Montague Strait has a broad, flat bottom with an equally steep ascent to Montague Island. This U-shape indicates that it was the work of a glacier which sculptured Montague Strait, as do the submerged hanging valleys, for example, Hanning Bay and Macleod Harbor, on the southwestern side of Montague Island, which are each 132 feet deep at their mouths and hang 528 and 330 feet, respectively, above Montague Strait.

The central outlet was between Hinchinbrook and Montague Islands. Hinchinbrook Island, which has also been called Nutcheck Island, is separated from Montague Island by a strait 7 to 10 miles wide, called Hinchinbrook Entrance. It has also been called Meiklejohn Entrance. The strait has a length of about twelve miles and is 900 to 1300 feet deep.[1] Its cross-section is a typical glacial U, for in the narrowest part there is an abrupt descent of 900 feet within ½ to ¾ of a mile of each shore and a flat, somewhat-basined bottom five miles wide. The lower slopes of the land are similar to the slopes below tidewater, for example, the precipitous character of Bear Cape on the eastern side of Hinchinbrook Entrance north of Port Etches, which is interpreted as a result of glacial oversteepening.

There are submerged hanging valleys on each side of Hinchinbrook Entrance. On the west are Rocky and Zaikof Bays, with maximum depths of 342 and 324 feet, submerged hanging valleys whose discordant junction with Hinchinbrook Passage, amounts to from 600 to 700 feet.

Port Etches east of Hinchinbrook Entrance also shows a discordant relationship, having a submerged hanging valley lip, with a descent of 318 feet to the bottom of Hinchinbrook Entrance, the water increasing in depth from 306 to 624 feet in less than one and one-half miles.

[1] See U. S. Coast and Geod. Survey Charts, Nos. 8550, 8515, and 8520.

At the lighthouse east of Hinchinbrook Entrance the glaciated ledges 185 feet above sea level bear striae which trend nearly due east. It is clear that these glacial scratches have been made by ice which came from Hinchinbrook Entrance of Prince William Sound, rather than from Orca Inlet to the northeast.

The granite bowlders in the glacial till here and the striations show, not only that the Prince William Sound Glacier extended out into the Pacific Ocean, but that it spread out in a bulb as soon as it was released from the confining walls of Hinchinbrook Entrance. This explains the eastward trend of the striae at Cape Hinchinbrook, in contrast with the southward trend in Hinchinbrook Entrance, and suggests that the Pacific side of Montague Island may have similar striae. A search for striae on Seal Rocks, which rise to a height of thirty-seven feet off the mouth of Hinchinbrook Entrance would be of great interest, both as to whether the Prince William Sound Glacier extended this far into the Pacific Ocean, and, if so, as to the direction of the ice movement here.

The easternmost outlet of Prince William Sound lies between Hinchinbrook and Hawkins Islands near the mouth of Orca Inlet. This outlet seems to have been of decidedly minor importance, so little ice moving through it that it was not eroded below sea level, excepting in the narrow Hawkins Island Cut-off and the still smaller Canoe Passage. There was, however, strong ice movement out of the northeastern portion of Orca Inlet.

The Driftless Area on Hinchinbrook Island. At the time of maximum glaciation, the Prince William Sound Glacier did not completely cover Hinchinbrook Island, which rose through it as a nunatak. The glacial deposits, at least in the southwestern portion, are entirely confined to its lower slopes, a large part of the island forming a driftless area.

This was determined in 1910 by observations at two points near the southwestern end of the island. Between English Bay, a cove on the southern side of Port Etches, and the top of Signal Mountain, a painstaking examination revealed glacial deposits only on the lower slopes. The till contained bowlders of granite and other foreign material not present in the country rock of Hinchinbrook Island, showing that the ice which carried this glacial drift was not a local glacier but the great ice sheet from northern Prince William Sound. The glacial deposits were found upon the gentler lower slopes, with a thickness of 15 or 20 feet, as on the headland southwest of English Bay; and a steepened slope, extending up to an elevation of less than 400 feet, showed the maximum thickness of the glacier. At about 400 feet there is also a small hanging valley east of English Bay.

At an elevation of 440 feet glacial deposits were entirely lacking, as we found by excavating through the thick peaty soil and residual deposits to bedrock. Frequent ledges from this point to the top of Signal Mountain revealed absolutely no rounding or striation, and there were no rounded pebbles excepting those traceable to certain, visibly-disintegrating, conglomerate ledges. The broad, rounded top of Signal Mountain, 1546 feet high, showed clear evidence of never having been glaciated, a careful examination revealing no erratics, no striae, and no rounded pebbles except those from the weathered ledges of conglomerate.

The second set of observations as to the height of glaciation was made at Cape Hinchinbrook on the Pacific side of the island near the new lighthouse on the southwestern tip. The lowest slopes were without glacial deposits, because they had been cut by the waves into great, precipitous seacliffs. Back of the cliffs, however, glacial till was ob-

served up to an elevation of 200 feet and glacial lakes (Pl. CLXXXII) nearly up to 400 feet, above which the absence of erratics showed that the area has not been ice-covered. The highest glaciated rock ledges seen were at an elevation of 185 feet, and these were strongly polished and striated.

Montague Island may also be partly driftless, though not for so large an area as Hinchinbrook Island, which had fewer local glaciers.

The rolling topography of the plateau-like southwestern end of Hinchinbrook Island continues throughout most of the southern half of the island. From sea level in Prince William Sound and on the Pacific Ocean it closely simulates a glacially-smoothed topography (Pl. CLXXXIII); but it is absolutely certain that it has never been covered by ice. It seems to be a part of the warped, southward-tilted peneplain, described by Spencer [1] and by Grant and Higgins,[2] and to owe its rounded topography to sub-aërial denudation. In it a few cirques of small local glaciers are cut.

The Glaciation of the Continental Shelf. In the Gulf of Alaska opposite the mouth of Prince William Sound there is an exceptionally wide continental shelf. Mention has already been made of the fact it is necessary to go 40 to 70 miles from shore before a depth of 600 feet is reached and that in the 56 miles between Hinchinbrook Entrance and Middleton Island the depth of water is only 200 to 450 feet.[3]

The question naturally arises as to whether glaciation has played any part in producing this broad, shallow continental shelf. The shelf, itself, is, of course, a feature which existed before and is independent of glaciation. That it may have been glacially modified, however, is suggested by (1) observations at Hinchinbrook lighthouse and at the southern end of Elrington Island which prove that the glacier from Prince William Sound once extended out into the the Pacific Ocean; (2) the deposits upon Middleton Island, which Dawson [4] has interpreted as glacial till. This suggests that the glacier moved outward across the continental shelf to or beyond Middleton Island, and that the Gulf of Alaska once had a great, tidal ice front similar to the barriers of the Antarctic region, but not floating because the water was too shallow.

It is difficult to determine whether the expanded glaciers actually pushed out to sea beyond the islands, glaciating the continental shelf. Aside from the glaciated condition of Hinchinbrook Entrance, there are three facts suggesting that they did. In the first place there is no visible deposit which can be considered representative of the terminal accumulation of this long period of expanded piedmont glacier in Prince William Sound. The inference seems warranted that it lies beneath the sea on the continental shelf. A second fact, suggesting the correctness of this conclusion, is the shallowing of the water south of the entrance to the sound. Soundings are not numerous enough to warrant definite conclusions on this point, but beyond the deep entrance to the sound between Montague and Hinchinbrook Islands, there is a broad area of shallower water with soundings of from 120 to 240 feet and beyond that only slightly deeper water. More detailed soundings are necessary before the significance of this can be definitely determined. A third noteworthy point is the depth of the entrance between the two islands. Its depth

[1] Spencer, A. C., Bull. Geol. Soc. Amer., Vol. XIV, 1903, pp. 117–132.

[2] Grant, U. S. and Higgins, D. F., Bull. 443, U. S. Geol. Survey, 1910, p. 15.

[3] See Chart 8502, U. S. Coast and Geod. Survey.

[4] Dawson, G. M., Notes on the Geology of Middleton Island, Alaska, Bull. Geol. Soc. Amer., Vol. 4, 1893, pp. 427–431; Brooks, A. H., Bull. 542, U. S. Geol. Survey, 1913, p. 43.

harmonizes with the theory that the Prince William Sound Glacier spread out into a great bulb on emerging into the sea and that it found freest passage and moved fastest in the narrow strait between Hinchinbrook and Montague Island and accordingly scoured out the passage through which it flowed. It could, of course, have produced the great depth in Hinchinbrook Entrance without extending beyond the islands, provided the ice moved through this passage long enough, or in a shorter period if the ice moved some distance out to sea.

It is not absolutely clear, however, that this has taken place and that the continental shelf is glaciated. Our doubt is based, chiefly, upon the low limit of glaciation at Hinchinbrook Island. If the ice extended only 400 feet above sea level and sloped between 25 and 60 feet to the mile, as in Prince William Sound, it could not well have advanced fifty-six miles farther to Middleton Island.

Another objection is found in the fact that the nature of the material from Middleton Island, which Dawson interprets as till, is such that it might equally well be an uplifted marine clay or silt, sprinkled with sub-angular, striated stones dropped by floating icebergs from glacier fronts in Hinchinbrook Entrance and the other outlets of Prince William Sound. The presence of crushed shells in the Middleton Island deposit and the fact of known, post-glacial uplift a short distance to the east on Wingham Island [1] support this interpretation.

The present conclusion therefore, regarding the glaciation of the broad continental shelf off Prince William Sound is that it is not certain whether the shelf has been actually overridden by an ice sheet or not, but that it has certainly been greatly modified through the former presence of glaciers, at least at the mouths of the various outlets of Prince William Sound. In any event, there are vast glacial accumulations on the continental shelf, made either by the deposition of till or of great volumes of glacial silt, supplied by rivers from beneath the melting ice, and distributed by the tidal currents, as at the Copper River delta today, with a scattering of glacial bowlders carried by icebergs. This silt-and-iceberg-bowlder combination is hard to distinguish from glacial till, or bowlder clay, unless marine shells are buried in it.

The Old Rock Floor of Prince William Sound. In various parts of Prince William Sound there are rock terraces at elevations of from 40 to 100 or 200 feet above sea level. These were first noted by Schrader.[2] In 1910 we saw others, which lead us to believe that these are remnants of the old rock floor of Prince William Sound. Our view is that there was no preglacial arm of the sea on the present site of Prince William Sound, but that glacial erosion has excavated the whole sound on the site of a preglacial lowland, whose rock floor was slightly above the present sea level. This lowland may have been crossed by preglacial stream courses, along the lines of which the fiords and the deeper submarine channels were excavated much more deeply than the intermediate areas.

This view is supported by the character of the low, flat-topped, rock terraces observed. On Culross Island near Passage Canal, for example, there is a low terrace which disappears to the eastward near Prince William Sound and to the westward near Culross

[1] Martin, G. C., Bull, 335, U. S. Geol. Survey, 1908, pp. 46, 64.

[2] Schrader, F. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, p. 404; Schrader and Spencer, House Doc. 546, 56th Congress, 2nd session, Washington, 1901, pp. 75–76.

Gilbert, G. K., Harriman Alaska Expedition, Vol. III, 1904, p. 176; Grant, U. S. and Higgins, D. F., Bull. 443, U. S. Geol. Survey, 1910, pp. 16-17.

Passage, where glacial erosion would have been most effective. The low rock benches on both sides of Port Nellie Juan near the mouth are not completely flat-topped, but have gently-undulating surfaces, as if modified by glaciation. The lack of seacliffs and talus slopes at the backs of these terraces suggests that they are of preglacial origin and not uplifted, wave-cut, post-glacial benches. There are two groups of flat-topped islets, the first between Montague and Green Islands, the second, called Porpoise Rocks, in Port Etches at the west end of Hinchinbrook Island. These are in places where remnants of the rock floor of Prince William Sound might well be preserved, because they are out of the paths of the most rapidly moving ice currents of the former piedmont glacier of Prince William Sound. The borders of each islet are cut into precipitous cliffs, and, as they are located where wave-work adequate for such erosion is now impossible, these cliffs are ascribed to waves from the iceberg-discharging front of the retreating Prince William Sound Glacier.

Recent Changes of Level in Prince William Sound. Mention has already been made of the post-glacial uplift on the continental shelf near Prince William Sound, and also a short distance to the east on Wingham Island, and of the shore forms near Columbia Glacier which Grant and Higgins [1] ascribe to recent uplift, but which we believe to be due to the work of iceberg waves. The latest change of level of the land adjacent to Prince William Sound, however, is a very slight downward movement.

The evidence that the latest change of level was a sinking of the land has been brought forward by Vancouver,[2] Hayes,[3] Schrader [4] and by Grant and Higgins.[5] Our own additional observations, made in 1910, are that there are dead trees in place and evidently killed by post-glacial submergence in three other localities. These are (1) on Point Countess, Flemming Island, on the southwestern side of lower Knight Island Passage, (2) on the southeastern side of Green Island, and (3) at Graveyard Point on the northwestern side of Montague Island. At each of these places the submergence may be still going on.

These localities are 6 to 32 miles apart and are 50 and 80 miles west of the areas of recent sinking on the Copper River delta and at Bering Lake, showing that the downward movement is widespread. On the other hand, it is extremely localized, for the areas of dead trees are very small and there is undisturbed, continuous forest close to sea level between these points of submergence, proving that there is no continuous sinking of the land. There is no reason for suspecting that this recent submergence in Prince William Sound is related to glaciation, as it probably is on the Copper River delta where isostasy and the loading of the seacoast with the outwash deposits of the Copper River glaciers may possibly explain the submergence.

Glacial Deposits. Most of the glacial deposits in Prince William Sound are submarine. In discussing the several fiord regions we have already described small areas of ground moraine, terminal and lateral moraine, outwash gravels, and deltas on the land, as well as the moraine bars and other deposits which mark terminal moraines in the sea. There

[1] Bull. 443, U. S. Geol. Survey, 1910, p. 18.

[2] See footnote on p. 278 in H. H. Bancroft's History of Alaska, San Francisco, 1886.

[3] Hayes, C. W., Nat. Geog. Mag., Vol. IV, 1892, p. 136.

[4] Schrader, F. C., 20th Ann. Rept., U. S. Geol. Survey, Part VII, 1900, p. 404.

[5] Grant, U. S. and Higgins, D. F., Bull. 443, U. S. Geol. Survey, 1910, pp. 17–18.

are scattered, small deposits of till and stratified deposits throughout Prince William Sound. Several typical ones are described below.

There is a flat plain of outwash gravels extending westward from Eyak Lake to Orca Inlet. One knob of morainic material rises through it. In the railway cuts near Cordova station of the Copper River and Northwestern Railway, there are fine exposures of the thick stratified gravels, interbedded with lake clays. In the town of Cordova, above the level of this outwash plain, there are deposits of stony till and gravel of considerable thickness. The mantle of ground moraine near Cordova probably extends higher toward the head of the fiord in the mountains, and lower near Prince William Sound and the Pacific.

We have no information regarding the extent of glacial deposits on Hawkins Island, but the relationships of the driftless area on Hinchinbrook Island suggest strongly that parts of this island, also, were not covered at the stage of maximum glaciation. The eastern end, lying in the jaws of two fiords, Orca Bay and Orca Inlet, where great glaciers formerly extended out from the Chugach Mountains, has probably been glaciated to a much greater height than Hinchinbrook Island. The glacial deposits on Hawkins Island, which has no igneous rock in place, include bowlders of coarse granite observed by Schrader and Spencer.[1] The geological map by Grant and Higgins[2] shows that these must have come from the Chugach Mountains to the northward between Port Gravina and Sheep Bay.

The low forelands on the northern sides of Hinchinbrook and Montague Islands have been glaciated, and, near sea level, bear thick glacial deposits, including both till and outwash, where we have observed them. The peninsula south of Port Fidalgo has a thick, flat-topped terrace of outwash gravels.

An exposure on Latouche Island, seen by us in 1910, was remarkable for the thorough cementation of the till and gravel by iron and copper-bearing waters. This is about 100 feet above sea level along the tramway between the steamer landing and the Bonanza copper mine.

Comparison of Prince William Sound and Yakutat Bay

Introduction. The glaciated portion of the Alaskan coast is divisible into three quite distinct physiographic provinces, all notably mountainous, and all supporting numerous glaciers, large and small. The southeasternmost of these is the irregular, fiorded coast below Cross Sound; the northwesternmost, which includes the region west of Controller Bay, is also an irregular, fiorded coast; and the coast between these is a straight, mountain coast, backed by a continuous, lofty mountain range. Yakutat Bay lies near the center of the latter, and Prince William Sound near the eastern end of the second province. The phenomena of present day glaciers and of former glaciation in these three sections present differences of moment and doubtless, ultimately, one of the important results of Alaskan glacier studies will be a consideration and interpretation of these differences. Although a beginning in such a comparative study might even now be made, on the basis of previous work by Russell, Reid, Gilbert, the United States Geological Survey, and others, we will not at the present time attempt it, but will confine ourselves to a

[1] Schrader, F. C. and Spencer, A. C., Geology and Mineral Resources of a Portion of the Copper River District, Alaska, House Doc. 546, 56th Congress, 2nd Session, Washington, 1901, p. 81.

[2] Pl. II, Bull. 443, U. S. Geol. Survey, 1910.

brief discussion of the similarities and contrasts of the two fields in which our own most detailed studies have been conducted,—Yakutat Bay and Prince William Sound.

Difference in Extent of Present Glaciers. We have no adequate records of precipitation for giving any basis whatsoever of a discussion of the comparative amount of snow that falls in these two areas, but there are some general considerations which lead us to the belief that there is less supply for glaciers in the Prince William Sound region than in that of the coast further southeast. In the first place, although glaciers are numerous and some of them large in the Prince William Sound region, they do not attain the magnitude of the glaciers along the straight coast which Yakutat Bay indents. Nor are there any of the piedmont glaciers and piedmont bulbs which characterize the central region from Controller Bay to Cross Sound and include such great ice bulbs as those of Bering, Malaspina, and La Perouse Glaciers. Another indication of the comparative weakness of the Prince William Sound type of glacier is the fact that there is not the same proportion of tidal glaciers as in Yakutat Bay. There are more tidal glaciers by actual count, but they are distributed over a far greater stretch of coast line. Furthermore, these tidal glaciers lie far back from the coast of the open ocean nearer the heart of the lofty mountains from which they are fed. The tidal glaciers are less active than those of Yakutat Bay. The great Columbia Glacier, for example, which is comparable to the Hubbard of Yakutat Bay, is far less active in iceberg discharge, and Columbia Bay is never so clogged with floating ice as to make navigation in small boats as difficult as in Yakutat Bay. Another point of contrast is the fact that neither the tidal glaciers nor those ending on the land in Prince William Sound appear to be parts of great through glacier systems comparable in size with those of the St. Elias and Fairweather regions. Although the valleys are ice-filled, and there are glacier passes, there does not seem to be the same ice-drowned topography that characterizes the region further southeast.

These differences in the extent of present glaciation of the two regions is apparently due primarily to two significant facts. The first of these is the position and nature of the mountains. In the Yakutat Bay region the lofty mountains lie along the coast, whereas in the Prince William Sound region the coastal mountains are lower and the main range of lofty mountains lies farther inland. The Chugach Mountains are of sufficient height to give rise to many and large glaciers, but not to a glacier system such as the massive, broad, and elevated St. Elias-Fairweather Ranges nourish. The second reason for the difference is in exposure to moisture-laden winds. In both cases the prevailing winds from the ocean reach the glacier-feeding mountains, but in the St. Elias-Fairweather mountains they reach the lofty chain without any intermediate barrier. To reach the mountains from which the Prince William Sound glaciers are fed, however, the west and southwest winds must first pass over a mountainous foreland on which a part of their vapor is necessarily condensed. Enough vapor still passes over this barrier and the glacier-supplying mountains are lofty enough to give rise to many large glaciers, but not to push them out to the mountain base, nor even to cause the tidal glaciers to discharge with the activity of those of Yakutat Bay.

Contrast in Size and Form of the Inlets. In a comparison of the glaciation of Yakutat Bay and Prince William Sound, and especially in considering the condition of former glaciation, it must be borne in mind that the two inlets are notably different in both size and form. As to size, Prince William Sound is several times as large as Yakutat Bay. In each case the form is divisible into two parts, a broad outer portion, and narrow inner

fiords. The inner fiord portions of the two inlets present no fundamentally important difference excepting that while the fiord portion of the Yakutat Bay inlet is confined to a single fiord only moderately branched, the corresponding portions of Prince William Sound are numerous and some of the fiord fingers are much branched. Aside from this difference the fiords of the two regions are so similar that they can be characterized in the same general terms, as narrow, mountain-walled fiords, with steep sides, truncated spurs, hanging valleys, and every indication of powerful glacial erosion as the chief agent in their shaping. They resemble also the fiords of the Inside Passage in southeastern Alaska.

Here, however, the resemblance between Yakutat Bay and Prince William Sound ends. In the former the single, moderately-branching fiord grades into a broadly-open, outer bay whose margins are formed not by mountains, but by a foreland of outwash gravel on one side and an ice plateau with a fringe of gravels on the other. But in Prince William Sound a dozen fiords from north, east, and west terminate in a roughly rectangular inlet, mountain-walled on three sides and on the other, or ocean side, defended from the ocean waves by mountainous islands.

Extent of Former Glaciation. At an earlier period, certainly at least several centuries ago, Yakutat Bay was completely filled with ice and a great Yakutat Bay Glacier discharged icebergs directly into the Pacific. From the evidence of the great amount of erosion performed and of the extensive deposits made it is clear that this period of extensive glaciation was of long duration. Prince William Sound, also, at an earlier period, and presumably at the same period as in Yakutat Bay, was occupied by an extensive glacier.

During the period of greatest glaciation, the branching tributary fiords of Prince William Sound were well filled with great glaciers and these, converging from all sides, coalesced in the broad sound to form an expanded ice sheet of the piedmont type, (Pl. CLXXXIV) which crowded up against the islands which lie across the mouth of the sound, crept out of the openings between them, and rose above the tops of all but the westernmost islands.

Origin of the Two Inlets. The fiord portions of both inlets as already intimated, are interpreted as the product of profound glacial erosion. The phenomena of glacial erosion are clear and convincing in Prince William Sound, as they are in the fiords of the Yakutat Bay inlet. Long-continued, vigorous ice motion along these fiords alone seems capable of explaining these phenomena.

With the outer bays the case is quite different. Outer Yakutat Bay, though doubtless modified in the northern portion by glacial erosion, and in the outer portion by glacial deposition, has not, in the main, the features which result from shaping by glacial erosion. We believe that this outer bay is a part of the shallow ocean bed on the continental shelf, protected by glacial occupation and deepened slightly by glacial erosion while the border of glacial outwash gravels was being laid down on the eastern side. What will be the result when the Malaspina Glacier disappears from the western side cannot be stated, but it seems probable that a part of its bed will become a bay of similar origin.

The origin of the main Prince William Sound is wholly different from this. It is a rock-enclosed bay. We are not in a position to state its origin with absolute certainty, though there are some facts of significance, pointing toward origin by glacial erosion. At first thought this may seem a very far-fetched hypothesis and one contrary to probabilities.

because of its great depth and breadth. Ice confined in narrow mountain valleys, and flowing through them with rapidity has quite generally been accredited with great erosive power; but ice which, emerging from narrow valleys, spreads out in a broadly-open depression, loses that velocity and concentration upon which effective ice erosion depends. While at first sight this seems to be a description of Prince William Sound, on closer study it is found to be not a strictly correct description. It is true that there are numerous, narrow, ice-eroded fiords, it is true that Prince William Sound is a broadly open depression, and it is true that there are great depths of water in it. But the apparent size of the depression is diminished by the fact that the western half is largely occupied by islands; and the apparent depth is less significant when it is noted that the deep parts are mainly, if not entirely, along rather narrow submarine channels (Fig. 72), in which it is a reasonable hypothesis to assume that the ice movement was more vigorous than elsewhere.

A fact of importance bearing on this problem is that the tributary valleys—the fiords—themselves have tributary hanging valleys at such elevations as to prove that before the ice invasion the main valley bottoms were well above the present sea level. This being the case, it is inconceivable that a few miles down their courses such valleys should become so deep as the bottom of Prince William Sound, for this would require an increase in valley depth of about 1500 feet in fifteen or twenty miles. Two hypotheses are possible to account for such valley discordance, (1) orographic changes, (2) differential glacial erosion. Of the former there is not only no evidence, but there are grave difficulties in accounting for the peculiar distribution of excessive depths, which harmonize with valleys and probable direction of ice motion and show no indication of origin by folding or faulting. In favor of glacial action is the fact that the ice was here for a long time, giving a cause known to exist, and the harmony of depths with lines of probable ice flow.

While we do not consider this explanation definitely established, we propose the following as an hypothesis for the conditions in the Prince William Sound region. Before the advent of the glaciers there existed here a mountain valley system tributary to the lowland of Prince William Sound, whose bottom was then at a level unknown at present but probably slightly above the present sea level. The glacial occupation gave rise to differential ice erosion, deepening and widening the fiords, excavating rock basins, and eroding to variable degree throughout the main Prince William Sound, but especially along the lines of most rapid flowage, namely out of the fiord mouth and down the middle of the eastern half of the sound. We believe, therefore, that Prince William Sound and the tributary fiords have been produced *without sinking of the land.*

Deposits of the Expanded Glaciers. In the Yakutat Bay region the expanded glaciers have left us well-marked records of both their outer stand and their recession. In our studies of Prince William Sound we have found no such distinct records. As stated above, the only interpretation that we can place upon the absence of terminal deposits is that these lie hidden from view in Prince William Sound and on the continental shelf south of the mouth of the sound.

Extensive recessional gravels, like those plainly visible in Russell Fiord, are not found in Prince William Sound. It is possible that this apparent absence of glacial deposits may be due to the fact that our studies here were not extensive enough, or that the forest covering has obscured the evidence, though this does not seem probable. If glacial de-

posits are lacking in Prince William Sound, then the recession of the expanded glaciers of Prince William Sound was different in nature from that of Yakutat Bay. It is possible that when the ice supply ceased to maintain the expanded piedmont glacier near its outer limit the ice margin was mainly in the waters of the sound and that the marginal deposits were made beneath the sea. This is a problem whose solution must depend upon further work, and it is possible that such work will discover marginal deposits in parts of the region which we did not visit.

Readvance of Glaciers. In Yakutat Bay there is clear evidence of a recent, but brief, great advance of the glaciers, during which they overrode but failed to completely remove a series of gravels of earlier date. Before this advance the glaciers had been farther back than now and forests spread above where the present glacier termini stand, as well as throughout the fiord where the overridden gravels lie. The recent history has been one of retreat and the forest has not yet advanced up to the glacier fronts, although formerly extending beyond them.

In Prince William Sound the glacier history has been notably different. Some of the glaciers are and have for some time been in recession, but even here there is no such extensive semi-barren zone between the glacier fronts and the forest as in Yakutat Bay. It is inferred, therefore, that where glaciers are retreating in the Prince William Sound region the recession has not been as rapid or as extensive as in Yakutat Bay. We have found no evidence in our study of Prince William Sound of an episode of later advance and retreat similar to that of Yakutat Bay, although we find a suggestion of one near by on the Copper River. Thus we conclude that, although in the period of greatest expansion of glaciers there was a probable synchronism of advance and retreat, since that time the two regions have had a quite different history. The glaciers of Yakutat Bay readvanced notably, but those of Prince William Sound apparently did not; the recent history of the Yakutat Bay glaciers has been one of rapid retreat, that of Prince William Sound has not, and, in at least one case, while general retreat was in progress in Yakutat Bay notable advance was in progress in Columbia Glacier and 14 other ice tongues of Prince William Sound. Ultimately these facts of difference in behavior of glaciers in different parts of the same coastal mountain region will doubtless have importance in interpretation of the causes of the difference; at present, with the limited range of our facts, we cannot present a verified interpretation. Whether it is due to local climatic variations, or to difference in time of response to more general climatic variations, or to influence of earthquake shaking, we cannot now determine.

Piedmont Glaciers and Ablation Moraines. Owing to the fact that large glaciers have extended with abundant supply from lofty mountains out upon lower, fairly-level land at their base, many of the glaciers of St. Elias-Fairweather mountains have spread out at the mountain base in piedmont bulbs and ice plateaus. This condition is characteristic of such glaciers in their lower ends in the Yakutat Bay region and its neighborhood to the northwest and southeast. It is entirely absent from the Prince William Sound regions, because one of the factors which give rise to it is absent, namely the termination of glaciers on lower, level land. Some of the glaciers end in the fiords, and these cannot, of course, assume the piedmont bulb form because expansion is checked by iceberg discharge due to the attack of sea water. Others, that end on the land, are confined in narrow valleys, and here too lateral expansion is prohibited. It is possible that some glaciers from side valleys entering larger main valleys may in this region expand into

small piedmont bulbs as they do in the Yakutat Bay region, but we have seen no examples of it. In the Copper River valley, however, there are some excellent examples, notably Miles and Allen Glaciers, which, descending from tributary valleys, expand on the broad floor of the Copper River canyon into true piedmont bulbs.

Among the glaciers that we have studied in Prince William Sound, only one or two bore an extensive burden of ablation moraine, such as one finds so commonly in the Yakutat Bay region. This may be due partly to the lesser friability of the rocks in the valleys through which the Prince William Sound glaciers flow, but it is partly due and perhaps mainly, to the fact that the glacier ends are not subjected to the long-continued wastage of the stagnant, piedmont condition. That the latter is the main reason is suggested by the fact that the piedmont bulbs of the Miles, Allen, and many other glaciers of the Copper River whose débris supply is from rocks of similar character to those of the Prince William Sound region are mantled with ablation moraine like that upon the glaciers of Yakutat Bay.

Causes of Recent Advances. As is fully stated in earlier chapters some of the glaciers of Yakutat Bay have undergone a remarkable, recent, spasmodic advance and transformation, which we believe to be the result of an impulse originating in sudden and unusual accession of snow in the glacier reservoirs during the severe shaking which accompanied the earthquakes of September, 1899. No similar recent advance and transformation was observed in the glaciers of Prince William Sound, but Childs Glacier on the lower Copper River had a similar episode in 1910.

The recent advance of the Columbia and 14 other Prince William Sound Glaciers is not in the same class with the advance of the 9 ice tongues of the Yakutat Bay region, either in rate of change or extent of the transformation. The recent advance of the Prince William Sound glaciers may have been a response merely to minor variations in climate; or it may be a response to some period of slight earthquake shaking in the Chugach Mountains. Upon this question it is at present impossible to speak with more definiteness, for we have a very limited series of observations of these glaciers. That the period of advance and recession, and that the extent and degree to which these variations take place, are different in the two regions is to be expected.

In order that the explanation of the variations may become established it is important that we have data covering a long period of years. At present our difficulty is the briefness of the period of observation, and the fact that we do not have a larger basis for comparison. When this work has covered a wider area and a greater number of years the Alaskan glaciers may be expected to give answers to some of the glacial problems not only of local, but of widespread and even general interest. Their activity and variability, their vastness and variety of form lead one to expect important results as an outcome of connected study of their characteristics and behavior.

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